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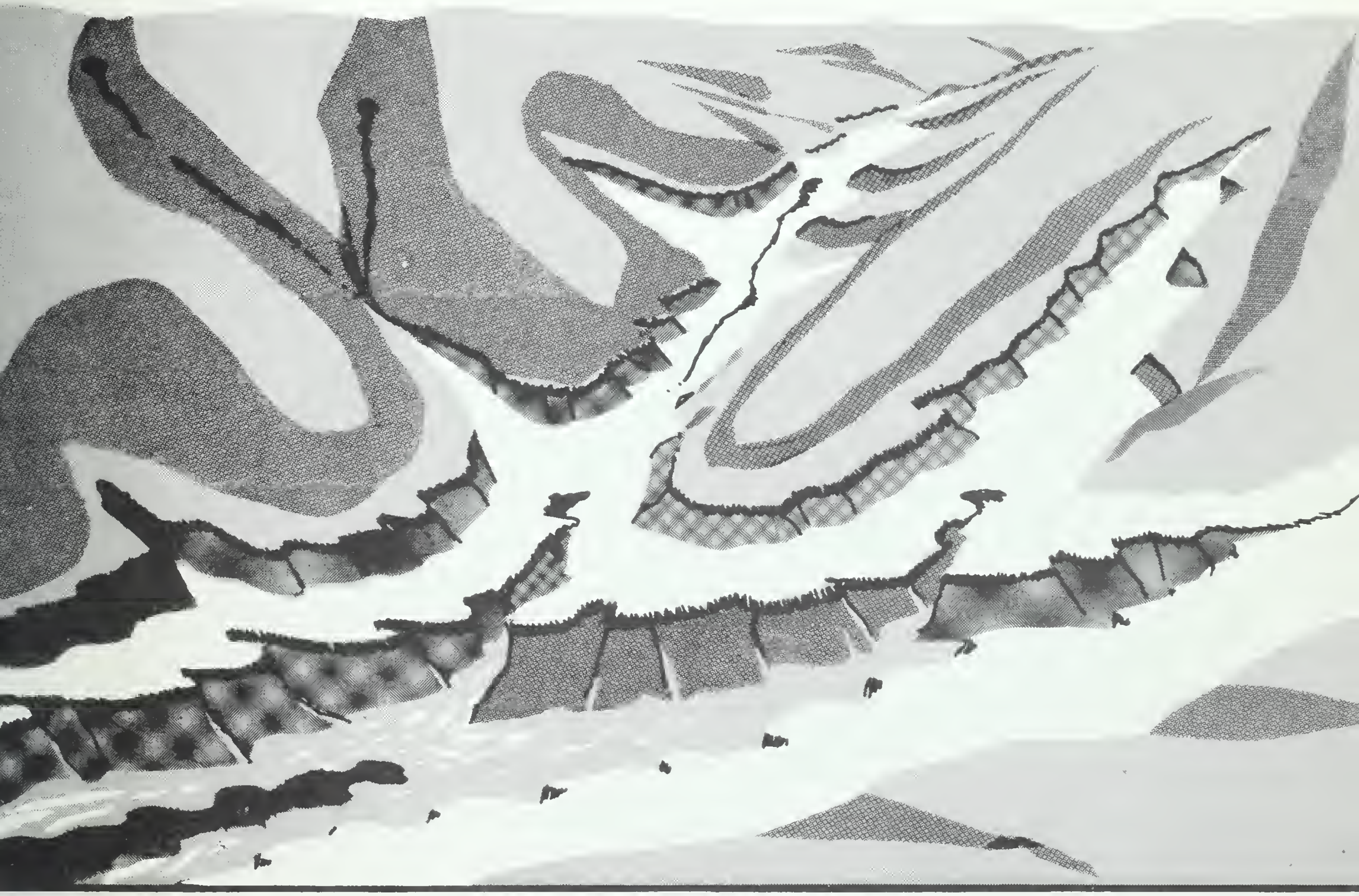

Gully Erosion

by

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PREFACE

This report was prepared under Bureau of Land Management Contract No. YA558-CT4-0011 to review the literature on incised channels, including gullies. Gullies are prominent features on many rangelands in the western United States as well as other locations throughout the world. Gully initiation, growth, and development are responsible for erosion of the landscape and downstream delivery of sediment.

Many land uses, including livestock grazing and surface mining, may influence gully erosion processes. Land managers often require information on the stages of gully evolution, current stability of gully systems, and estimates of long-term gully erosion rates and sediment yields. While there are very few standardized methods for evaluating gully systems, a great deal of information on gully erosion processes has been generated. The purpose of this report is to make information on gully erosion available to resource specialists and provide a conceptual framework to help evaluate gully systems and gully erosion processes.

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EXECUTIVE SUMMARY

1. Although gullies have been investigated in numerous locations around the world, general conclusions about their initiation, dynamics, evolution and control have rarely been forthcoming because climate, soils, and vegetational conditions have been very different (Schumm et al, 1984).

2. Gullies are one member of a continuum of incised channels and, therefore, studies of rills, arroyos and channelized streams can provide additional valuable information on gully processes. All incised channels must be regarded as components of an integrated fluvial system.

3. A gully may develop rapidly in a short time period due to the exceedance of an intrinsic or extrinsic geomorphic threshold.

4. The response of a system to gullying is complex. That is, secondary responses, which result from negative feedback mechanisms, complicate the adjustment of the system to change.

5. Thresholds at which gullies are initiated can be predicted, for essentially homogeneous soil and climatic conditions, by quantifying relative shear stress and stream power indices.

6. All incised channels follow essentially the same evolutionary trend with eventual development of relatively stable conditions which can be recognized and quantified. Stages of evolution include: initiation, headward migration, channel widening, channel slope reduction, reduction of bank angles, deposition of sediment and establishment of vegetation. Small gullies and arroyos may be filled eventually and obliterated. Large gullies and arroyos tend to form a new floodplain at a lower elevation than the pre-incision level.

7. The nature of the sediments eroded from, and transported through the incised system, significantly affects the morphology of the channel and the nature of the channel adjustment.

8. Assessment of the current stability of a gully system depends on the ability to identify the new equilibrium state of the channel. Empirical relations can be generated to quantify the equilibrium state. These relations depend on criteria such as: channel slope, valley slope, valley width, bank heights and angles, channel top widths, and discharge or drainage area data.

9. Prediction of long-term gully erosion rates and sediment yields is hampered by the stochastic nature of the distribution, in time and space, of precipitation events. Gully migration rates and ultimate gully configurations (i.e. onsite effects) can be estimated. However, the effects of gully erosion on sediment yields and sediment delivery ratios (i.e. offsite effects) are more difficult to predict. The mere presence of a gully does not necessarily imply high sediment yield.

10. Conceptual frameworks, that include data requirements and analyses, for the determination of both onsite and offsite effects of gullying are provided.

1. INTRODUCTION

Incised channels have lowered their beds by degradation, thereby setting in motion a period of considerable channel instability with the potential for serious erosion, damage to agricultural production, and downstream delivery of sediment. Incising channels also pose hazards to man-made structures, especially bridges, and a major cause of bridge failure is the deepening and widening of an incised channel (Shen and Schumm, 1981). However, the deeper more efficient channel is a desirable way to move floodwaters through a valley.

Incised channels range in size from small rills, most commonly observed on roadcuts, to major incised channels that may be on the order of 10 to 15 meters deep. Clearly, they are major sediment producers. For example, the infamous Rio Puerco arroyo, although comprising only 20 percent of the Rio Grande drainage basin above Elephant Butte Reservoir and producing only 8 percent of the runoff, delivers almost 50 percent of the sediment that reaches the reservoir (Patton, 1973).

The worldwide gully problem has been studied by numerous investigators, and general descriptions of gully morphology and evolution are available for gullies in many parts of the world, such as Israel (Seginer, 1966), New Zealand (Blong, 1970; Gair and Williams, 1964), Nigeria (Ologe, 1972), the mid-western U.S.A. (Bariss, 1971; Beer and Johnson, 1963; Bertrand and Woodburn, 1964; Daniels and Jordan, 1966; Ireland et al, 1939; Thompson, 1964; Piest and Spomer, 1968), and western U.S.A. (Heede, 1967, 1970, 1974; Schumm and Hadley, 1957; Brice, 1966; Cooke and Reeves, 1976; Antevs, 1952; Bryan, 1925, 1928a; Dodge, 1902; Duce, 1918; Hastings, 1959; Leopold and Miller, 1956; Miller and Wendorf, 1958; Rich, 1911; Swift, 1926; Thornthwaite et al, 1942; Tuan, 1966; Graf, 1979; Love, 1979).

The results of much of this research is disappointing because conclusions of a general nature are rarely forthcoming. For example, several attempts to measure rates of gully extension have been made, but results from different areas are rarely comparable because of differences in climate, soils and vegetation and because the most spectacular gullies have usually been studied rather than a representative sample of incised channels.

In the U.S.A., the problem of channel incision and extensive gully development was first appreciated in the agricultural areas of the East and Midwest, where the conversion from natural vegetative cover to crops of various types led to considerable erosion. The extreme gullying and loss of productive land as a result of cotton farming in the southeastern United States is an excellent example of problems faced by the early agriculturalists (Ireland et al, 1939). This problem has now been largely rectified because the uplands which were the areas most severely gullied, have been revegetated and stabilized by mechanical techniques and by land-use changes. However, a new problem has arisen. It relates to efforts to reduce flooding of valley-bottom agricultural land by the straightening and alignment (channelization) of existing stream channels. The channelization process steepens and straightens the streams, and therefore, erosion in the straightened channels is tremendously accelerated. The channel deepens and enlarges, and as a result of the incision the entire drainage basin may be rejuvenated if geologic controls are lacking.

The channelization problem is an order of magnitude greater than that associated with the upland gullies because channelization affects valley floors and potentially, the entire drainage network. Large incised channels may be stabilized by construction of drop structures and by bank stabilization procedures, but in effect, the base level of the entire drainage area has been lowered by channelization and the restoration of the original condition is probably impossible. In addition, the incision and enlargement of the channel produces large quantities of sediment that move downstream, where its deposition aggravates flood problems, reduces reservoir capacity, and induces channel instability.

It is apparent that incised channels cannot be considered in isolation from the remainder of the fluvial system. Therefore, there is a spatial aspect to the problem as well as a temporal one because channel morphology and dynamics evolve through time, and the focus of erosion and deposition shifts within the drainage network through time.

STATEMENT OF THE PROBLEM

In arid and semi-arid environments, active gully initiation, growth and development is responsible for significant erosion of the landscape, which in turn leads to significant downstream delivery of sediment. To date there are no commonly accepted techniques that quantify gully initiation, growth and development under varying conditions of land use. However, the mere presence of gullies does not necessarily indicate high sediment yields (Thornes, 1984). A generally utilizable technique is required to quantify the effects of gullying for land-use planning, environmental impact analysis, and reservoir sedimentation purposes.

OBJECTIVES OF THE REPORT

The primary objective of this report is to review the extensive literature on incised channels and to develop a conceptual framework that can be utilized in the formulation of an easily applied technique which can be used to analyze:

(1) Threshold conditions for gully initiation
(2) The stages of gully evolution
(3) Current stability of a gully system
(4) Long-term gully erosion rates and sediment yields

APPROACH

This report is based on the geomorphic, and engineering literature that deals with incised channels of all orders. The conclusions drawn from the surveyed literature are in general geomorphic in nature and, therefore, it is appropriate that a thorough understanding of geomorphic principles and fluvial system behavior is provided for the reader (Chapter 2). A discussion of drainage network rejuvenation is presented because the development of incised channels of all types can be considered to be an aspect of drainage-network adjustment (Chapter 3). Specific discussions of the various members of the incised channel continuum, gullies, arroyos and channelized streams, are provided in Chapters 4 and 5. The soil mechanics aspects of channel degradation, soil piping and bank instability are discussed in Chapter 6. Finally, the literature is synthesized to provide a summary of our knowledge of both the onsite and offsite effects of gullying (Chapter 7).

Geomorphologists recognize that the interface between the atmosphere and the solid earth, the landscape, is dynamic. Drainage basins and their components, slopes and channels, are either adjusting rapidly to altered conditions (instability) or they are in dynamic equilibrium with present conditions. Billions of dollars have been spent during the past century to insure that the landscape and its components are in a condition most favorable for man's activities. Sometimes the results have not been satisfactory and often this is because the geomorphic character of the landscape was neglected.

Incised channels have been a major land management and conservation problem at least partly because the different evolutional stages in the erosional development of these landforms have been ignored. For example, Heede (1974, p. 261) is of the opinion that "once we can quantitatively describe the gullies in terms of stage of development, decision making in land and water management will be substantially improved." and (1974, p. 271): "Watershed managers would have a useful tool if gully stage could be expressed in terms of erosion rates and sediment yields." In fact recognition of erosional and depositional stages of landform evolution is the basis of geomorphology.

In addition to the importance of the evolutionary concept, there are two ways in which an understanding of the geomorphology of a region can provide essential information upon which land use planning and conservation activities can be based. These are (1) the identification within a relatively stable landscape of potentially unstable locations where change is likely to occur, and (2) the recognition that conservation activities in unstable channels will be most successful when implemented at the proper evolutionary stage of development and at selected places, where the structure or the technique will act with the evolving system rather than in opposition to it.

To consider the first of the above geomorphic approaches, it is necessary only to recognize that many people view the landscape as being static and think that change is undesirable. However, Mosley (1980) has demonstrated that much high-country erosion in New Zealand is natural. This means that it will be very difficult to control and, in fact, cessation of this erosion may produce major channel adjustments downstream.

The construction of small erosion and water-control structures in the millions of square miles of western United States has for the most part, proceeded in response to local decisions, and therefore, the overall program has developed in a piecemeal fashion. In some areas a high percentage of the small check dams and gully plugs fail. This may be due to inadequate design and construction, but it may also be that the geomorphic situation is unfavorable. That is, within a region or even a watershed there are favorable and unfavorable conditions for control. These can be recognized by an application of geomorphic principles to land management.

Frequently, after years of failure to induce stability in unstable channels the efforts are finally successful. However, this success may be achieved only when the system has evolved to a suitable condition, and considerable effort and money could be saved if geomorphic criteria for recognition of landform susceptibility to control were available. For example, experimental and field studies show that deposition inevitably follows erosion in alluvial valleys. The recognition of this sequence in the field should enable proper timing and proper placement of the structures.

The following brief statements contain the essence of a geomorphic approach to land management and erosion control:

1. The land surface is complex, dynamic, and it changes with time.
2. The land surface may respond dramatically during a short time due to the exceeding of a geomorphic threshold (Schumm, 1973, 1977).
3. The response of a complex landform, for example a drainage basin, to change is itself complex. That is, secondary responses will complicate the adjustment of the system to change (Schumm, 1973, 1977).

The first statement requires little elaboration. There is a progressive erosional evolution of a landscape through time. The manner of change depends on geology, climate, vegetation and, of course, land use. In addition, the components of the system respond complexly to changed environmental conditions.

The second statement relates not only to dramatic erosional events, that can be directly related to destruction of vegetation, but also to the random occurrence of slope failure and channel trenching (arroyos, gullies) in response to storm events and land use. That is, within an area of otherwise similar conditions erosion occurs in a seemingly random fashion. The explanation frequently is that geomorphic threshold has been locally exceeded.

An example, which will aid in the understanding of the concept is based on studies of gully formation in semi-arid regions. Sediment delivered to the semi-arid valleys tends to be stored there progressively steepening the slope of the valley floor. This continues until flushing of the sediment takes place by gullying. When the gullies heal the process begins again (Schumm and Hadley, 1959; Schumm, 1973). This obviously is a long-term process, and an entire region will not be in the same stage of development. Therefore, during a major flood event only some valleys will trench. As will be demonstrated later in this report it is possible to identify the incipiently unstable valleys and to take steps aimed at preventive conservation.

The third concept, complex response, refers to the search of a channel for a new equilibrium position. For example, channel incision is normally followed by deposition and perhaps renewed incision. This complex response renders prediction difficult, but progressive change is not the normal course of events. In fact, in steep high-sediment producing watersheds episodic behavior may be normal. That is, periods of high sediment production and rapid channel incision may be followed by low sediment production and sediment storage. The effect of incision and its resulting complex response and episodic behavior are discussed in more detail later.

The above geomorphic principles may be, in fact, hypotheses that require further testing, but they provide a basis for preventive conservation and a means of evaluating the potential for success of conservation programs.

The basic geomorphic approach requires a proper perspective of the drainage system. It is necessary to step back from the immediate problem and to view it in the perspective of the fluvial system. Therefore, a systems approach is used in which gully or unstable channel reach is viewed as simply a component of a larger geomorphic system. In some cases, such an approach will reveal that the problem is really a local adjustment of a channel that is expectable, and is natural system behavior. Therefore, it is necessary to study both relatively stable and rapidly adjusting landforms and to document landform history.

It is clear that in many areas of landform and landscape management, tactics are used, and a strategy is absent. A strategy can be developed when the fluvial system is comprehended in both engineering and geomorphic ways. The historical approach of the geomorphologist can be combined with the engineer's quantitative description of the incised channel to provide a quantitative basis for prediction of channel behavior. The ability to predict will yield successful tactics that can be employed by the conservation engineer.

An additional example of the geomorphic approach is provided by the soil conservationist, A.B. Foster (1964). His discussion of gully stabilization indicates that it is necessary to establish a gradient or a grade at which a revegetated gully floor will be stable. He refers to this as a silting grade. He states that "due to the differences in soils, there is no exact grade that is stable under all conditions. Probably the best and easiest way to determine a silting grade is to investigate the grade of gullies in the area that appear to be stable - that is, that are not actively eroding in the bottom of the channel." Foster here is suggesting that the stable and unstable landforms be compared, and that, if an unstable landform is to become stable, it must assume the morphology of existing, stable landforms.

TERMINOLOGY

A major problem that is basic to discussion of incised channels is the definition of these features. The greatest confusion appears to be related to the definition of a gully, which obviously is a term central to any discussion of channel incision. For example, a very broad use of the term is its application to submarine, delta-front "gullies" of the Mississippi River delta (Roberts et al, 1980), which are the result of submarine slumping and mudflows.

In the glossary of the American Geological Institute (1972, p. 318), a gully is defined as "a) A very small valley, such as a small ravine in a cliff face, or a long, narrow hollow or channel worn in earth or unconsolidated material (as on a hillside) by running water and through which water runs only after a rain or the melting of ice or snow; it is smaller than a gulch. b) Any erosion channel so deep that it cannot be crossed by a wheeled vehicle or eliminated by plowing, esp. one excavated in soil on a bare slope."

Gregory and Walling (1973, p. 369-370) probably present the best statement concerning gullies and their characteristics when they write: "The general characteristics of gullies, implicit in the many alternative definitions, include the facts that they often have ephemeral streamflow, they are often incised into unconsolidated materials, and they may have a V-shaped cross section, where the subsoil is fine in texture and resistant to rapid cutting, but U-shaped in material like loess where the soil and subsoil are both equally susceptible to erosion. In size they are larger than rills, they are usually bordered by steep sides and heads which often have the appearance of erosional scarps, and they are usually so deep that restoration is impossible with normal tools, and they cannot be crossed by a wheeled vehicle or eliminated by ploughing."

The qualifications in this statement are a clear indication of a lack of a consistent definition. Part of the difficulty is that the term is taken from common usage, which was general, and a more specific meaning is now being imposed. For example, the term gully was first used for an erosional feature in 1657, and it is probably a corruption of the term gullet, which is Middle English for a "defile gully or ravine." (Little, 1964, p. 844).

The key factor that is missing from all of the definitions is that a gully is an unstable landform; it is dynamic and changing and it is part of a drainage network transformation.

Gullys can be classified by size for agricultural purposes (Bennett, 1939) as small, less than three feet deep; medium, three to fifteen feet deep; and large, more than fifteen feet deep, but size alone is no criterion. However, the distinction made by Daniels (1966, p. 51) is important. He distinguished between valley-side gullies and entrenched streams in western Iowa. He states that, "A valley-side gully is a small, steep-walled, sharply incised, elongate depression on valley sides." Whereas, an entrenched stream "flows in a steep-walled trench cut into alluvium." The implication is that the valley-side gully is the result of an expansion of the drainage network whereas the entrenched stream is the result of deep incision of a pre-existing channel. The emphasis on a steep-walled trench is important because it conveys the suggestion of a major change.

Therefore, in addition to the criterion of instability there is another important factor, significant alteration of the landform. That is, the gully forms where no channel or only an insignificant channel existed previously, and tne entrenched stream is the result of a major change or metamorphosis of an existing channel by natural or man-induced causes. Further, the major change will lead invariably, to future major adjustments of the landform and perhaps of the drainage system as a whole. In all cases, a major alteration of the drainage system is underway.

In order to provide a basis for further discussion the following incised-channel classification is proposed, which leads from the smallest to the largest channel (Table 1-1).

1. Rill - An ephemeral channel, which is one of the smallest channels formed by runoff and which can be destroyed by plowing or by frost action. They can be seasonal in nature and are a result of overland flow. The commonest are parallel rills formed on bare roadcuts (AGI, 1972, p. 538; Schumm, 1956).

2. Gully - A relatively deep, recently formed eroding channel that forms on valley sides and on valley floors where no well-defined channel previously existed. Two major gully types have been recognized: a) Valley-side gully which is an extension of the valley network and which is incising into soil, colluvium and perhaps bedrock. b) Valley-floor gully which may be discontinuous or continuous and which is incising into alluvium and perhaps bedrock.
3. Entrenched Stream - A deep trench resulting from incision of an existing channel or a man-modified stream. The best examples of the former are southwestern arroyos and of the latter, the channelized streams of the Midwest and Southeast.

All of these incised channels can exist within a drainage network. When a drainage network is rejuvenated, an entrenched stream forms in the main valleys, valley-floor and valley-side gullies add low-order channels to the network, and rills form on steep slopes (Table 1-1).

Table 1-1. Classification of Incised Channels

1. Rills
2. Gully
    a. Valley-side gully
    b. Valley-floor gully
3. Entrenched Stream
    a. Ephemeral or Intermittent - Natural (Arroyo)
    b. Perennial or Intermittent - (Channelized Stream)
4. Drainage System Rejuvenation

There is a need to discuss some further aspects of terminology relating to within-channel features, primarily the breaks in the longitudinal profile or nickpoints. Within incised channels and particularly at the head of an incising channel there is an abrupt break in the longitudinal profile of that channel. This break point or bend point is a knickpunkt. This German term has been corrupted, and frequently it is spelled knickpoint or nickpoint.

The nickpoint, as it will be referred to here, is a location on the profile where there is an abrupt change of elevation and slope. This feature is commonly referred to as a headcut, which migrates up the valley floor or along an existing channel. The nickpoint can be a point, when the break in gradient is a headcut or a channel scarp, as it is sometimes called, or the nickpoint can be a steeper zone if the headcut has rotated and extended itself along the longitudinal profile of the stream. In this case, the change in elevation is distributed over a channel reach. In this report the abrupt change in elevation at the head of a gully will be referred to as a primary nickpoint or headcut (Fig. 1-1). When the change occurs within an incising channel, it will be referred to as a secondary nickpoint. A steeper reach of channel, which represents a headward migrating zone of incision, will be referred to as a nickzone.

The terminology used in this report may replace some in wide use. However, it is hoped that it will not introduce additional confusion to what is a very complicated problem.

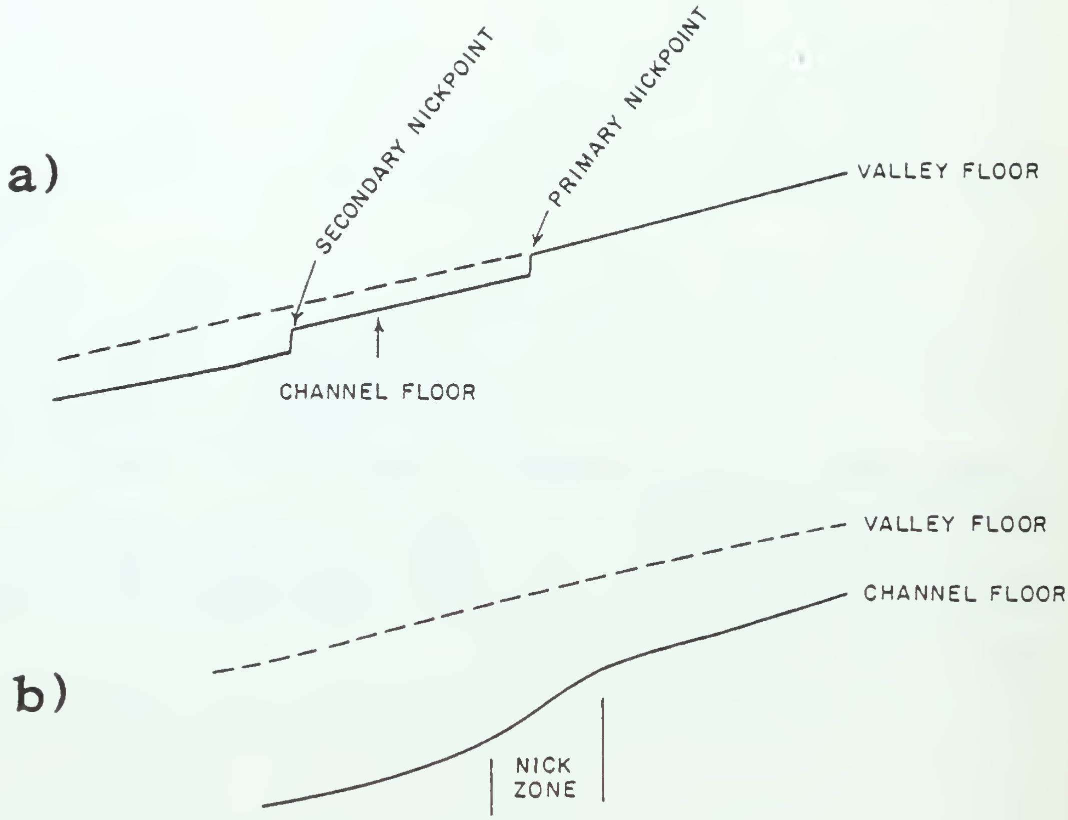

Figure 1-1. a) Sketch showing primary and secondary nickpoints.

b) Sketch showing a nickzone.

2. FLUVIAL SYSTEM BEHAVIOR

An understanding of the cause of incised channels and the effects of the incision is important for practical reasons. The causes of incision are well known, although the reason that a specific channel is incised may be impossible to determine. For example, considerable controversy surrounds the origin of southwestern incised channels (arroyos), which is indicated by the following plaintive comment by Cooke and Reeves (1976, p. 15): "A reader of the prodigious arroyo literature may be justifiably perplexed by the shifting current of conflicting arguments, the discharge of unsubstantiated assertions, the pools of controversy and the shoals of abandoned hypotheses."

CAUSE OF INCISION

A channel exists with its characteristic morphology because the eroding force exerted by concentrated flowing waters exceeds the resistance of the earth materials in which it is flowing. Therefore, all channels end before a divide is reached, as the drainage area above the channel head is unable to provide sufficient runoff to permit additional elongation of the channel. In addition, numerous valley floors are well vegetated and contain only very small channels or none at all. Any disruption of valley-floor vegetation or a significant increase of runoff or peak discharge can trigger major incision of the valley alluvium (Graf, 1979). The obvious causes of incised channel development are listed on Table 2-1.

The development of an incised channel at a given location may depend on controls acting at that site, or the upstream or downstream changes that affect the site. In addition, the response can be the result of changes imposed on the system, extrinsic controls, such as climatic fluctuations, change of land use, or channel modification. There are also intrinsic controls that are inherent in the system. Changes such as natural channel shift, steepening of valley floor by deposition and natural meander cutoffs all occur naturally. Yet, if they are of sufficient magnitude, they can initiate channel incision. Examples of

Table 2.1 Causes of Channel Incision

A. Decreased Erosional Resistance

1. Decreased vegetational cover owing to increased agricultural activity, urbanization, timbering, overgrazing, fire and droughts.
2. Surface disturbance causing decreased permeability and cohesion.

B. Increased Erosional Forces

1. Constriction of flow by dikes, vegetation, etc.
2. Concentration of flow by roads, trails and ditches.
3. Steepening of gradient and energy slope by deposition of sediment (valley plug), channelization and meander cutoffs, base-level lowering by meander cutoffs, shift of stream mouth, main channel incision, lowering of lake or reservoir level, lowering of discharge and flood peak.
4. Increased discharge and flood peaks.
5. Decrease of sediment load.

these influences will be considered later. Some of the principle causes of incised channels of various orders are given on Table 2-2. See also Cooke and Reeves (1976, p. 16) for a discussion of the causes of arroyo development in southwestern United States.

THRESHOLDS

A very important point is that channel incision can be inherent in the erosional development of a valley, and when geomorphic thresholds are exceeded incision will occur (Schumm, 1977). It is the identification of these thresholds that is an important objective of the geomorphologist.

Thresholds have been recognized in many fields. Perhaps the best known to geologists are the threshold velocities that are required to set in motion sediment particles of a given size. With a continuous increase in velocity, threshold velocities are encountered at which movement begins, and with a progressive decrease in velocity, threshold velocities are encountered at which movement ceases. Thresholds in hydraulics are described by the Froude and the Reynolds numbers, which define the conditions at which flow becomes supercritical and turbulent. Particularly dramatic are the changes in bedform characteristics at threshold values of stream power (Simons and Senturk, 1977).

In these examples, an external variable changes progressively, thereby triggering abrupt changes or failure within the affected system. This is an extrinsic threshold. That is, the threshold exists within the system, but it will not be crossed and change will not occur without the influence of an external variable.

Thresholds of the other type are intrinsic thresholds and changes occur without a change in an external variable. An example is long-term progressive weathering that reduces the strength of slope materials until eventually there is slope adjustment and mass movement (Kirkby, 1973), and in semi-arid regions sediment storage progressively increases the slope of the valley floor until failure occurs by gullying. Intrinsic thresholds are probably common in geologic systems, but only geomorphic examples will be considered here.

Table 2-2. Causes of Incision, Five Types of Incised Channels

1. Rills - Decreased erosional resistance by decrease of vegetational cover and surface disturbance.
   Increased erosive force by artificial steepening of slope (road cut, tailings pile).

2. Valley-side Gullies - Decreased erosional resistance by decrease of vegetative cover and surface disturbance.
   Increased erosive forces by concentration of runoff by roads, trails and ditches, and mass movement and piping.
   Base-level lowering and rejuvenation of main channel or drainage network.
   Increased runoff and flood peaks.

3. Valley-floor Gullies - Decreased erosional resistance by decreased vegetative cover and surface disturbance.
   Increased erosive force by constriction of flow, steepening of valley-floor gradient by sediment deposition and by base-level lowering.
   Increased runoff and flood peaks and decreased sediment load.

4. Entrenched Channel - Increased erosive forces by narrowing of channel. Steepening of gradient by channelization or base-level lowering.
   Increased runoff and flood peaks.
   Decrease of sediment load.

5. Drainage Network Rejuvenation - Increased erosive forces by base-level lowering.
   Increased runoff and flood peaks.
   Decreased sediment load.

The idea that landform change can occur without a change of external controls challenges the well-established basis of geomorphology, that landform change is the result of some climatic, base-level of land-use change. Therefore, the significance of the geomorphic threshold concept is that abrupt erosional and depositional changes can be inherent in the normal development of a landscape and that a change of an external variable is not always required for a geomorphic threshold to be exceeded and for a significant geomorphic event to ensue (e.g., gullying).

There are also extrinsic geomorphic thresholds. For example, in common usage "thresholds" can be the result of either cause or effect. That is, we speak of hydraulic, velocity, shear and stream power thresholds above which sediment moves or banks fail, but we can also speak of bank, channel and slope stability thresholds, when the forces causing the failures are not clearly identified and understood. Therefore, geomorphic thresholds can be of two types, and they can be defined in the following way. A geomorphic threshold is a threshold of landform stability that is exceeded either by intrinsic change of the landform itself, which may involve a change of strength of the materials involved, or by a change of an external variable.

Recent field and experimental work support the concept of geomorphic thresholds. The best examples result from investigations of gully distribution and stream pattern variability. Field studies in valleys of Wyoming, Colorado, New Mexico, and Arizona revealed that discontinuous gullies, can be related to the local oversteepening of the valley-floor surface (Schumm and Hadley, 1959). For example, the beginning of gully erosion in these valleys tends to be localized on steeper convex reaches of the valley floor. Carrying this one step further, for a given region of uniform geology, land use and climate, a critical valley slope will exist above which the steepest reach of the valley floor is unstable. In order to test this hypothesis, measurements of valley-floor gradient were made in the Piceance Creek basin of western Colorado (Patton and Schumm, 1975). The area is underlain by oil shale, and the potential environmental problems that will be associated with the development of this resource are considerable.

Within the Piceance Creek area, valleys were selected in which discontinuous gullies were present. The drainage area above each gully, was measured on maps, and valley slopes at the point of gully development were surveyed in the field. No hydrologic records exist, so drainage basin area was selected as a substitute for runoff and flood discharge. When this valley slope is plotted against drainage area, the relationship is inverse (Fig. 2-1), with gentler slopes being characteristic of large drainage areas. As a basis for comparison, similar measurements were made in ungullied valleys, and these data are also plotted on Figure 2-1. The lower range of critical slopes of the unstable valleys coincide with the higher range of slopes of the stable valleys. In other words, for a given drainage area it is possible to define a critical valley slope above which the valley floor is unstable.

Note that the relationship (Fig. 2-1) does not pertain to drainage basins smaller than about seven square miles. In these small basins, variations in vegetative cover, which are perhaps related to the aspect of the drainage basin or to variations in the properties of the alluvium, prevent recognition of a critical threshold slope. Above four square miles there are only two cases of stable valley floors that plot above the threshold line, and one may conclude that these valleys are incipiently unstable and that eventually a flood will cause erosion and trenching in these valleys.

Using Figure 2-1, one may define the threshold slope above which trenching or valley instability will take place in the Piceance Creek area. This has obvious implications for land management if the slope at which valleys are incipiently unstable can be determined. Corrective measures can be taken to artificially stabilize such critical reaches, as they are identified.

Note also that both intrinsic and extrinsic threshold conditions can be recognized on Figure 2-1. The steepening of the valley slopes by deposition will result in the exceeding of an intrinsic geomorphic threshold, whereas an increase of runoff would have the effect of shifting the points to the right until some cross the threshold line. Obviously, drainage area cannot be increased but, if discharge were

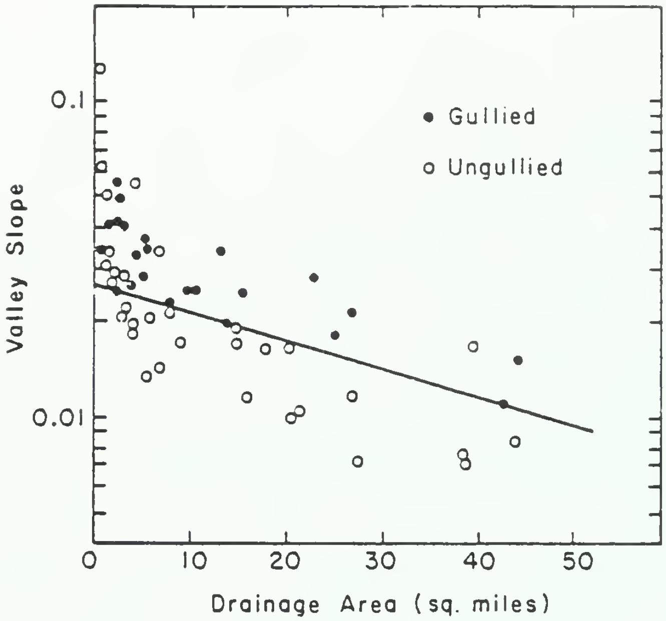

Figure 2-1. Relation between valley slope and drainage area, Piceance Creek Basin, Colorado. Line defines threshold slope that separates gullied from ungullied valley floor. (Patton and Schumm, 1975)

plotted on the abscissa of Figure 2-1, the points would shift to the right with an increase of runoff.

The concept of thresholds, as applied to alluvial deposits in the western U.S., is illustrated by Figure 2-2, where the decreasing stability of an alluvial fill is represented by a line indicating increase of valley slope with time. Of course, a similar relation would pertain if, with constant slope, sediment loads decrease slowly with time. Superimposed on the ascending line of increasing slope are vertical lines showing the variations of valley-floor stability caused by flood events of different magnitudes. The effect of even large events is minor until the stability of the deposit has been so reduced by steepening of the valley gradient that during one major storm, erosion begins at time A. It is important to note that the large event is only the most apparent cause of failure, as it would have occurred at time B in any case.

Studies of alluvial deposits in drylands and steeplands suggest that large infrequent storms can be significant but only when a geomorphic threshold has been exceeded. It is for this reason that high-magnitude, low-frequency events may have only minor and local effects on a landscape. This conclusion has bearing on the work of Wolman and Miller (1960) concerning the geomorphic importance of events of large magnitude. They concluded that, although a major amount of work is done by events of moderate magnitude and relatively frequent occurrence, nevertheless, the large storm or flood may have a major role in landscape modification.

These and other observations indicate that a major event may be of either major or minor importance in landscape modification, and an explanation of the conflicting evidence requires further consideration of the threshold concept. Some landscapes or components of a landscape have apparently evolved to a condition of geomorphic instability and these landforms will be significantly modified by a large infrequent event whereas others will be unaffected. Therefore, there will be, even within the same region, different responses to the same conditions of stress.

When some landscape components fail by erosion whereas others do not, it is clear, that erosional thresholds have been exceeded locally.

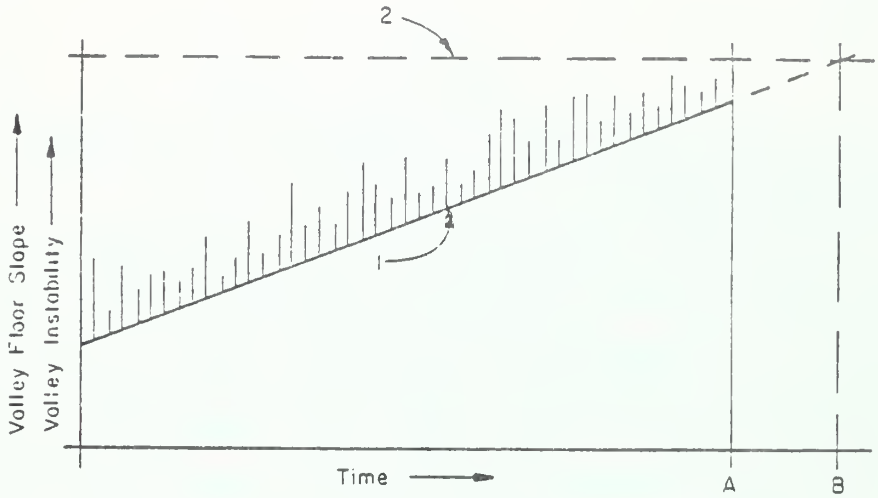

Figure 2-2. Hypothetical relation between valley-floor gradient and valley-floor instability with time. Superimposed on line 1, representing an increase of valley-floor slope, are vertical lines representing instability of the valley floor as related to flood events. When the ascending line of valley floor slope intersects line 2, representing the maximum slope at which the valley is stable, failure or trenching of the valley alluvium will occur at time B. However, failure may occur earlier, at time A as the result of a major storm or flood event. (Schumm and Hadley, 1958)

An important task for the geomorphologist is to locate incipiently unstable components of a landscape. The recognition of geomorphic thresholds within a given region will be a significant contribution to the understanding of the details of regional morphology as well as providing criteria for the identification of incipiently unstable landforms and a basis for stabilization or modification of these critical landforms.

EFFECTS OF INCISION

Channel incision can be considered to have onsite, upstream and downstream effects.

Onsite effects appear to be obvious, as a channel incises to a new base level. However, experimental studies and the examination of the details of channel incision in several field areas reveal that the response of a channel to incision may be complex, with deposition following incision (Schumm, 1973, 1977, 1980). In fact, the channel incision may be episodic, as the channel is periodically overwhelmed by sediment (Womack and Schumm, 1977; Schumm, 1977). Although this response may appear to make prediction of channel response very difficult, it may, in fact, provide a means of recognizing when a channel is approaching a new measure of stability.

In areas of high sediment production the storage and flushing of sediments results in episodic channel behavior and variations from deposition to erosion in relatively short periods of time. Thus, although a channel may be incising, it will cycle through periods of aggradation and degradation as pulses of sediment move through the channel.

Obviously, channel incision will be propagated upstream. The incision of a channel effectively lowers the base level of the upstream area and causes the upstream migration of nickpoints, which rejuvenate tributaries and which may produce valley-side gullies unless the system heals itself or unless the rejuvenation is stopped by artificial means.

The downstream effects of channel incision are initially obvious. The rejuvenation of tributaries and the considerable bank erosion accompanying channel incision produces large quantities of sediment that in turn may produce downstream aggradation. Hence, various components of the fluvial system can be out-of-phase, with gullying and channel incision taking place at one location and with deposition and channel filling and valley-plug development elsewhere. Excellent examples of such fluvial system behavior are provided by Happ et al (1940) and Trimble (1974).

COMPLEX RESPONSE

A landscape component is not always in a condition of grade, balance or equilibrium. The existence of the threshold suggests an inability of the landform to adjust readily to a new equilibrium condition. During experimentation, a miniature (15 m by 10 m) drainage system (Schumm and Parker, 1973; Schumm, 1977) was rejuvenated by a slight (10 cm) change of base level. As anticipated, base-level lowering caused incision of the main channel and development of a terrace (Fig. 2-3 a, b). Incision occurred first at the mouth of the system, and then progressively upstream, successively rejuvenating tributaries and scouring alluvium previously deposited in the valley (Fig. 2-3 b). As erosion progressed upstream, the main channel became a conveyor of upstream sediment in increasing quantities, and the inevitable result was that aggradation occurred in the newly cut channel (Fig. 2-3 c). However, as the tributaries eventually became adjusted to the new base level, sediment loads decreased, and a new phase of channel erosion occurred (Fig. 2-3 d). Thus, initial channel incision and terrace formation was followed by deposition of an alluvial fill, channel braiding, and lateral erosion, and then, as the drainage system achieved stability, renewed incision formed a low alluvial terrace. This low surface formed as a result of the decreased sediment loads, when the braided channel was converted into a better defined channel of lower width-depth ratio.

The experimental results indicate that an event causing channel incision within a drainage basin (tilting, changes of base level, climatic and/or land use) automatically creates a negative feedback (high sediment production) which results in deposition; this is eventually followed by incision of alluvial deposits as sediment loads decrease.

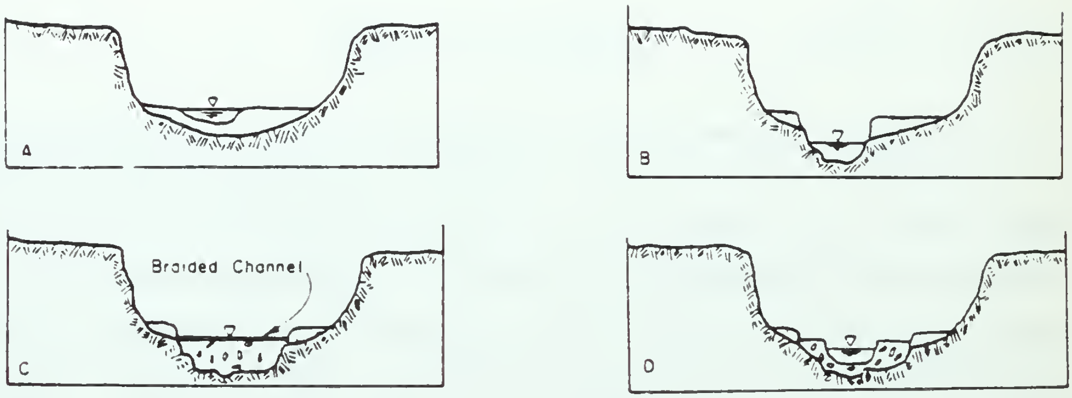

Figure 2-3. Diagrammatic cross sections of experimental channel 1.5 m from outlet of drainage system (base level) showing response of channel to one lowering of base level.

A. Valley and alluvium which was deposited during previous run, before base lowering. The low width-depth channel flows on alluvium

B. After base level lowering of 10 cm, channel incises into alluvium and bedrock floor of valley to form a terrace.

C. Following incision, bank erosion widens channel and partially destroys terrace. An inset alluvial fill is deposited, as the sediment discharge from upstream increases. The high width-depth ratio channel is braided and unstable.

D. A second terrace is formed as the channel incises slightly and assumes a low width-depth ratio in response to reduced sediment load. With time, in nature, channel migration will destroy part of the lower terrace, and a flood plain will form at a lower level. (Schumm and Parker, 1973)

The complex response observed during the experiments is apparently a quest for a new equilibrium by a complex system. In fact, the changes documented in Figure 2-3 are minor as compared to those accompanying major periods of aggradation or degradation which result from major climatic or base-level changes. Under these circumstances, the response of the system is episodic.

EPISODIC BEHAVIOR

It is well established in the geomorphic literature that a major reduction of base level will cause progressive downcutting and readjustment of the stream gradients until a new graded or equilibrium situation has been developed. In fact, the fluvial system may not be capable of degrading in this way when sediment movement is great (Hey, 1979).

Where major incision has occurred in alluvium and relatively weak rocks, evidence of pauses in the erosional downcutting are found, but this is usually attributed to some external influence, such as variations in climate the rate of base-level change, or variations in the rates of uplift of the sediment source area. However, recent studies in the Douglas Creek drainage basin of western Colorado support the idea of discontinuous downcutting (Womack and Schumm, 1977). The investigation of recent erosional history of this valley shows that modern incision of the valley fill began after 1882. Yet, unpaired, discontinuous terraces formed during downcutting. Elsewhere, these have usually been explained by the shifting of an incising channel laterally across the valley floor (Davis, 1902). In the Douglas Creek Valley of western Colorado, however, downcutting was discontinuous. In fact, during pauses in downcutting there was deposition (Fig. 2-4). That is, during incision of the main channel, there was rejuvenation of tributaries and a progressive increase in sediment yield from upstream. Sediment loads became so great that downcutting ceased and deposition began. Deposition continued until it was possible for the channel to incise again and to continue the downcutting process.

The Douglas Creek situation appears to conform to the observations of Born and Ritter (1970) who mapped six discontinuous and unpaired

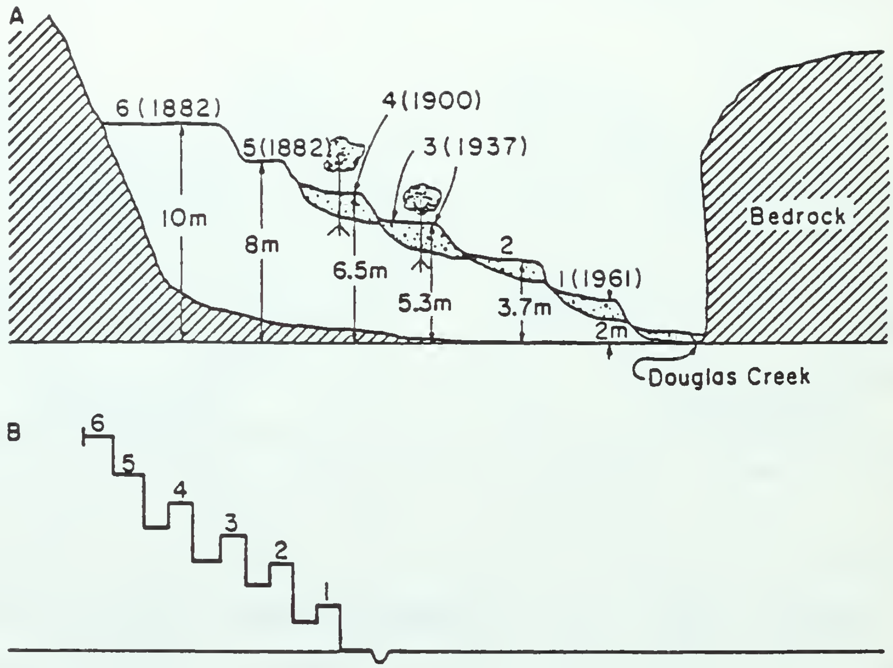

Figure 2-4. A. Sketch of Douglas Creek valley showing erosion surfaces formed since 1882. Age of surfaces is based on tree-ring dating and historical data. Note burial of trees by deposition. Surfaces 5 and 6 were present before modern erosion began after 1882.

B. Summary of behavior of Douglas Creek. Vertical segments indicate incision or deposition, horizontal segments periods of relative instability. (Womack and Schumm, 1977) terraces at the mouth of Truckee River where it enters Pyramid Lake in Nevada. A reduction of the water level in Pyramid Lake reduced the base level of the lower Truckee River, but instead of simple downcutting commensurate with the lowering of the base level, the channel in fact, paused as many as six times. A similar response has been described during erosion of hydraulic mining debris in valleys of the Sierra Nevadas, California (Wildman, 1981) and during late-Wisconsin incision of the Little Sioux River (Hoyer, 1980a, 1980b).

Sediment production from a basin will vary greatly, as sediment is stored and remobilized. This was observed during an experimental study of drainage-network rejuvenation (Schumm and Parker, 1973; Schumm, 1977).

The episodic behavior, as identified in Douglas Creek and during some experiments obviously can have significant effects on channel stability, sediment yields and can explain the apparent unpredictability of channel behavior as well as explaining bridge and bank failure under presumably safe hydraulic conditions.

3. DRAINAGE NETWORK REJUVENATION

The development of incised channels of all types can be considered to be an aspect of drainage-network adjustment. The present drainage pattern is assumed to reflect modern climatic and hydrologic conditions, but in many areas the channel network does not completely fill the valley network, and it is capable of expansion if erosional conditions change (Fig. 3-1). Therefore, an understanding of drainage network growth and evolution will provide a basis for further consideration of specific types of incised channels.

Bennett (1939, p. 605) provides an example of network expansion on an Alabama farm. Initially there were three channels with a few short tributaries on 153 acres, but after 100 years of cultivation the three channels had expanded to fill the area by development of more than a thousand tributary gullies. The farm became a maze of gullies and channels as a result of the removal of vegetation and the resulting extension of the previously poorly-defined drainage networks. What occurred on the farm is what Strahler (1956) referred to as "a drainage basin transformation" (Fig. 3-1), in which low-order streams extend headward with the addition of tributaries to develop a higher-order drainage network with a higher drainage density (total stream length divided by drainage area). This can take place as a result of agricultural activities, or during geologic time, as a result of climatic change or base-level lowering. Therefore, study of natural drainage network growth and evolution may be revealing in terms of incised-channel development.

The first detailed analysis of drainage-network growth was qualitative. Glock (1931) selected topographic maps of drainage basins in various stages of erosional development and arranged them in a sequence to show growth of the drainage patterns. The assumption is made, of course, that the drainage pattern is initiated on an essentially smooth plane due to uplift or retreat of the sea.

Glock (1931) described the phases of development as follows: initiation, the beginnings of the pattern and the development of a shallow skeletal drainage pattern on the undissected area; elongation,

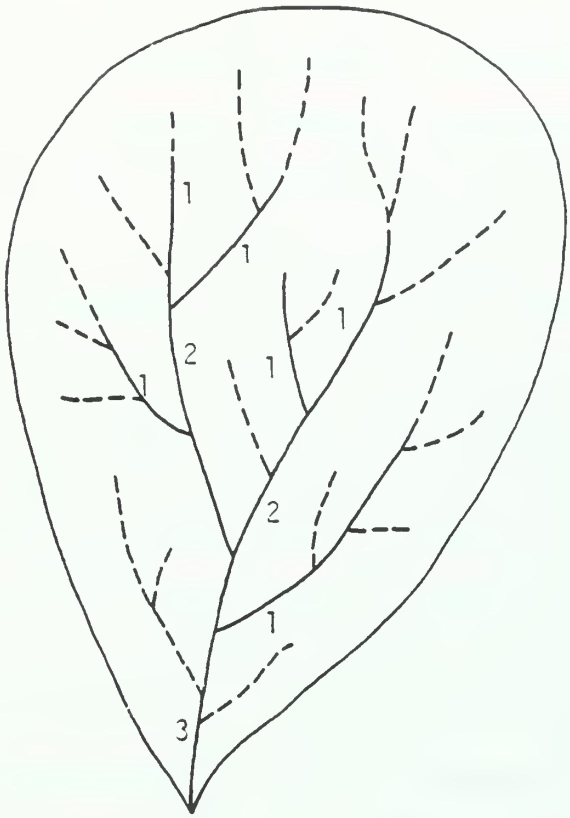

Figure 3-1. Third-order drainage network showing present channels (solid lines) and new channels (dasned lines) following rejuvenation. After rejuvenation the network becomes a fourth order drainage system. Numbers are Strahler-order of present channels.

headward growth of the main streams; elaboration, the filling in of the previously undissected areas by small tributaries; maximum extension, the maximum development of the drainage pattern; and abstraction, the loss of tributaries as relief is reduced through time. The period of abstraction persists until the number of channels on the surface is reduced to a minimum.

The evolutionary development of a drainage network can also be studied experimentally. This has the advantage of controlled conditions, and the elimination of variables that complicate the process. Three experimental studies of pattern development performed in the Rainfall Erosion Facility (REF) at Colorado State University will be reviewed in some detail in the remainder of this Chapter. The first experiment dealt mainly with the development of parallel drainage lines, rills, on an essentially planar surface, whereas the other experiments were concerned with dendritic drainage patterns.

The REF is a 9 by 15 meter container, that was filled with a mixture of sand, silt and clay. The surface could be graded to any desired configuration. The experiment began with the application of artificial precipitation from sprinkler heads to the sediment surface. The development of channels under these conditions was documented by photographs and measurements. Adjustments in the evolving drainage network were produced by changing precipitation intensity and duration or by base-level lowering.

RILLS

An investigation of the effect of slope inclination and shape on rills was performed by Mosley (1972, 1974). He applied artificial precipitation at an intensity of 90 mm per hour during 9 experimental runs in the REF.

In addition to the morphologic description of the rills, sediment and water discharge were measured as they were delivered from the rills.

According to Mosley (1972, p. 40) after the surface was graded and shaped to the desired configuration, a 10 cm section of the lowermost partition was removed to create a vertical face. This base-level lowering established the depth to which the rills could incise. As water was applied to the surface by artificial precipitation, at first the water infiltrated into the soil, but after two or three minutes runoff commenced. Instead of flowing downslope as a thin sheet, the water collected almost immediately into more or less distinct flow lines leading to the heads of the rills, which were slowly cutting headward. The results obtained by Mosley show the effect of the slope and shape of the surface plane, concave (converging), and convex (diverging) on the morphology of the erosion patterns and on the sediment produced by the eroding area. Gradients of the plane surface ranged from 2 to 11%. Unfortunately, the 11% slope could not be maintained throughout the REF and only 8.5 meters of the total length of the container was used. In addition, the length of the concave and convex surfaces are different. Nevertheless, Mosley's results reveal the effect of surface slope and shape on the erosional process.

Examination of Mosley's results show that patterns developed on the planar surfaces of different inclination. As expected, the length of channels increased on the steeper slopes. The undissected area between the rill heads and the top of the REF became smaller with increased slope as the rills lengthened. This is the so called "belt of no erosion" (Horton, 1945). Erosion was occurring in this zone, but it was not channel erosion. Sheet wash and raindrop impact were dominant.

Drainage density (total length of channels divided by drainage area) increased with increased slope for all surface configurations and, of course, a very different drainage pattern developed on the three different surfaces. At slopes of about .07 the drainage density of the concave surface was higher than that of either the planar or convex surface but at the steepest slopes (>0.1) there was little difference between the drainage densities of the planar, convex or concave surfaces (Mosley, 1974). This is because for the sediment used and the precipitation applied the maximum possible drainage density was achieved (73 m/m^2).

The drainage patterns change both with inclination and shape of the surface. On the planar surface, there is an elongation of the drainage patterns and an increased parallelism. If the width-length ratio of the rill basins is used as an index of shape and only the longest drainage systems are measured, the ratio decreases from 0.25 to 0.17 as the slope steepens. Of course, on the concave surface the flow converges to produce a more highly branching network in contrast to the diverging flow on the convex surface. The inclination and shape also significantly affects erosion rates. Mosley demonstrated excellent relations between drainage area and sediment yields, and between sediment yield and slope (Figs. 3-2, 3-3).

The steeper and converging flow patterns produce more sediment, although runoff per unit area was the same from all surfaces (800 ml/min).

Mosley's work provides an interesting example of drainage network extension into essentially planar, concave, and convex areas which explains the differences in sediment yield and pattern development in such morphologically different areas. In addition, it suggests that the design of artificial landforms, particularly in relation to disposal of spoil and tailings should consider not only the longitudinal profile of the slope, but the transverse configuration with a convex upward transverse profile being the least erodible because of the dispersion rather than concentration of the runoff.

DENDRITIC PATTERN DEVELOPMENT

The dendritic drainage pattern is the normal pattern that results from channel erosion in the absence of structural controls such as faults, folds, or joints. It is the extension or transformation of a dendritic drainage network that produces gullies and advances the network into as yet undissected headwater areas. Some problems of channel incision, therefore, are associated with the headward extension of the drainage pattern and the response of the main channel to the increased sediment production and flood peaks that result from an extended and better-integrated channel network.

The REF was used in another experimental study that followed the growth of a dendritic drainage pattern following base-level change

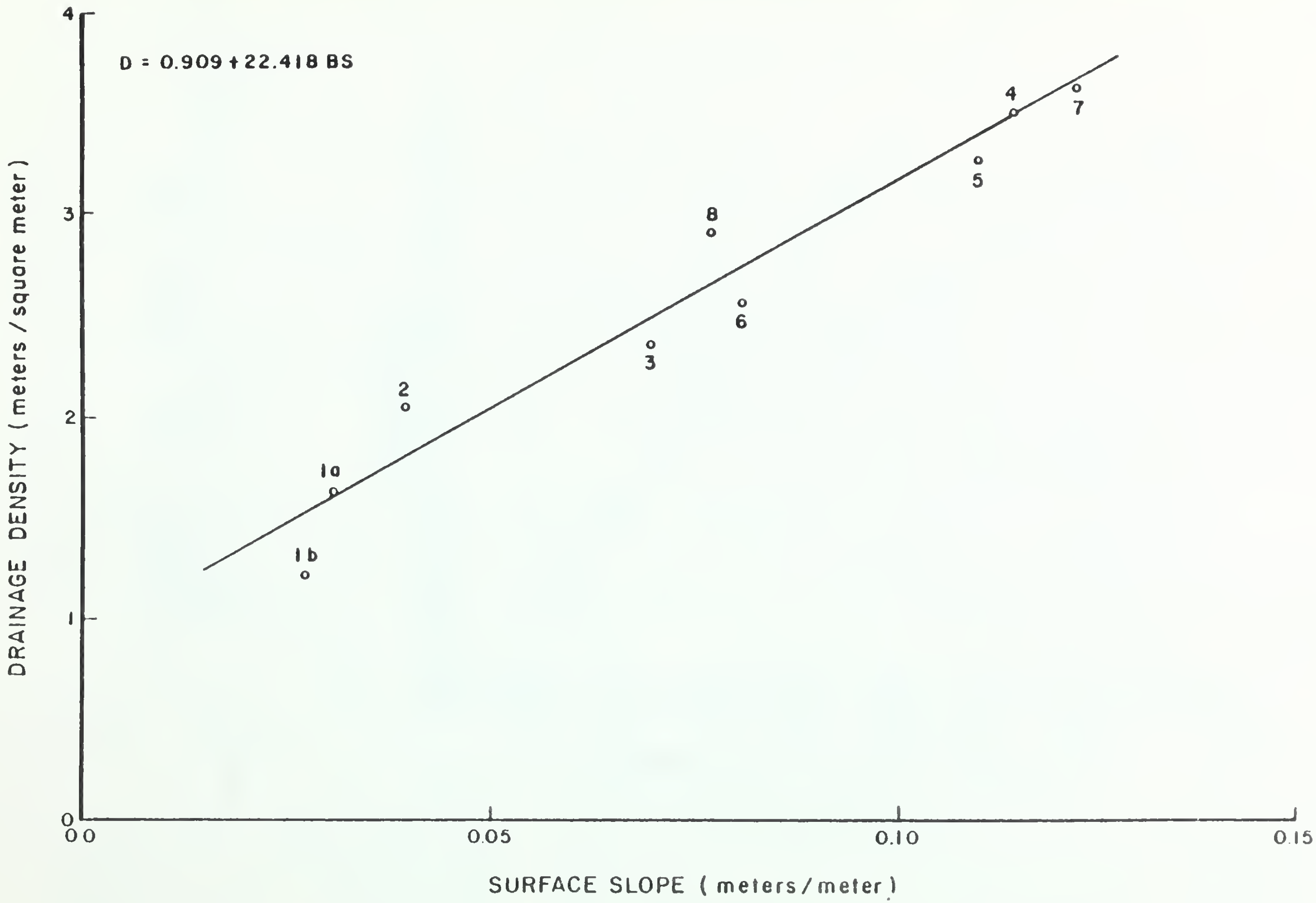

Figure 3-2. Mean drainage density as a function of surface slope, surfaces 1 to 8. Numbers identify each surface. (Mosley, 1974)

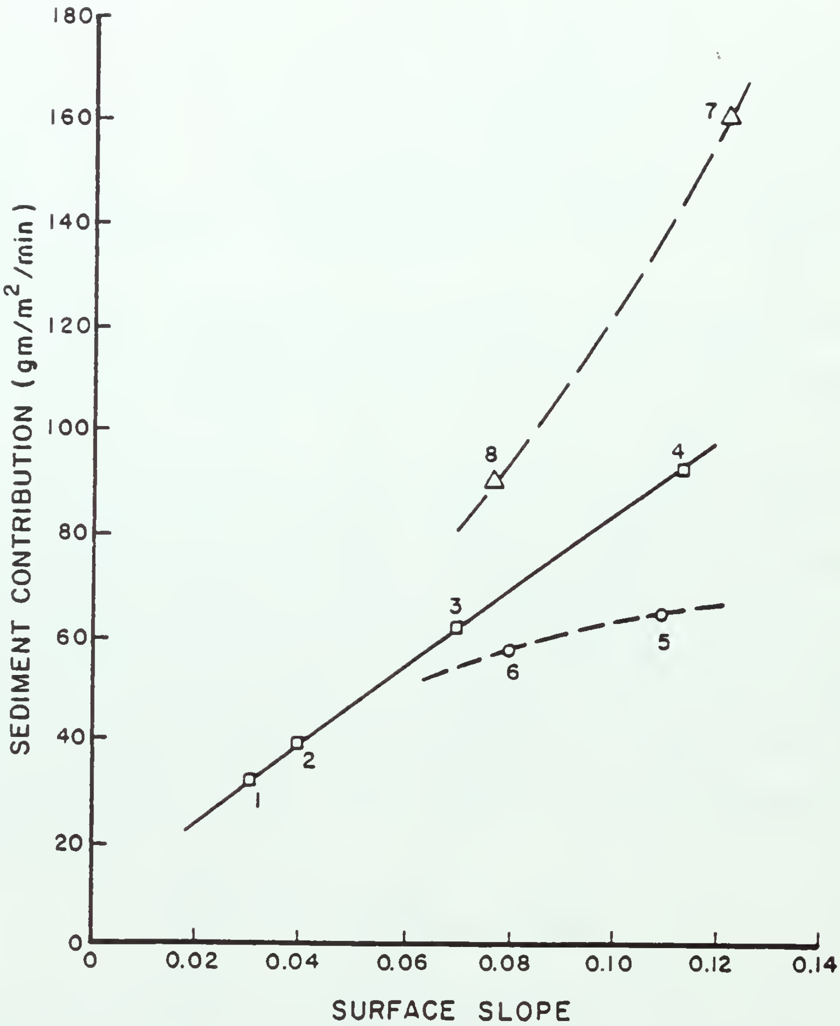

Figure 3-3. Mean sediment yield as a function of surface slope. Points 7 and 8 represent a converging surface; Points 1 to 4 represent a planar surface, Points 5 and 6 represent a diverging surface. (Mosley, 1974)

(McLane, 1978). Unlike the previous experiment, the surface of the sediment in the REF was graded with an inclination of 0.0075 toward the center of the container and toward the outlet (Fig. 3-4). The area of this model drainage basin was 106 m^2. Note that in spite of the effort to form two intersecting planes, the contours show some irregularity and especially some depressions that concentrated the flow.

The first phase of the experiment was to document the initial channel-network development on an undissected surface. Phase two of the study involved rejuvenation of the basin by a lowering of base level and documentation of the resulting expansion of the existing channel system.

McLane's (1978) observations, as the network developed, are important with regard to the controls on development of drainage networks and the paths followed by drainage patterns. He notes that the surface configuration of the model basin was designed to direct runoff toward the longitudinal centerline of the basin, where it would collect and flow to the basin outlet. Channel erosion progressed rapidly headward along this centerline to form a wide, main channel with vertical banks. Water was concentrated in depressions on the surface, and points at which interconnected pond systems drained over the banks became sites of tributary nickpoint initiation. As these nickpoints migrated headward into the pond systems, major tributary channels were formed.

Two conclusions may be derived from the preceeding description. First, the channel formation proceeds along lines of greatest water supply. Channels grew headward along the path of the rudimentary surface drainage system, and they bifurcated in response to the form of these pathways. Headward growth of a drainage network is hydrophilic, advancing always toward maximum water supply (Schumm, 1956, p. 620).

Secondly, and more fundamentally, the drainage system is a product of the topography on which it forms. Initial topography shapes the rudimentary drainage system which, in turn, provides the paths of maximum water supply, which is followed by the headward developing incised channels. Thus, initial micro-topography controls the formation of the final network pattern in the model basin.

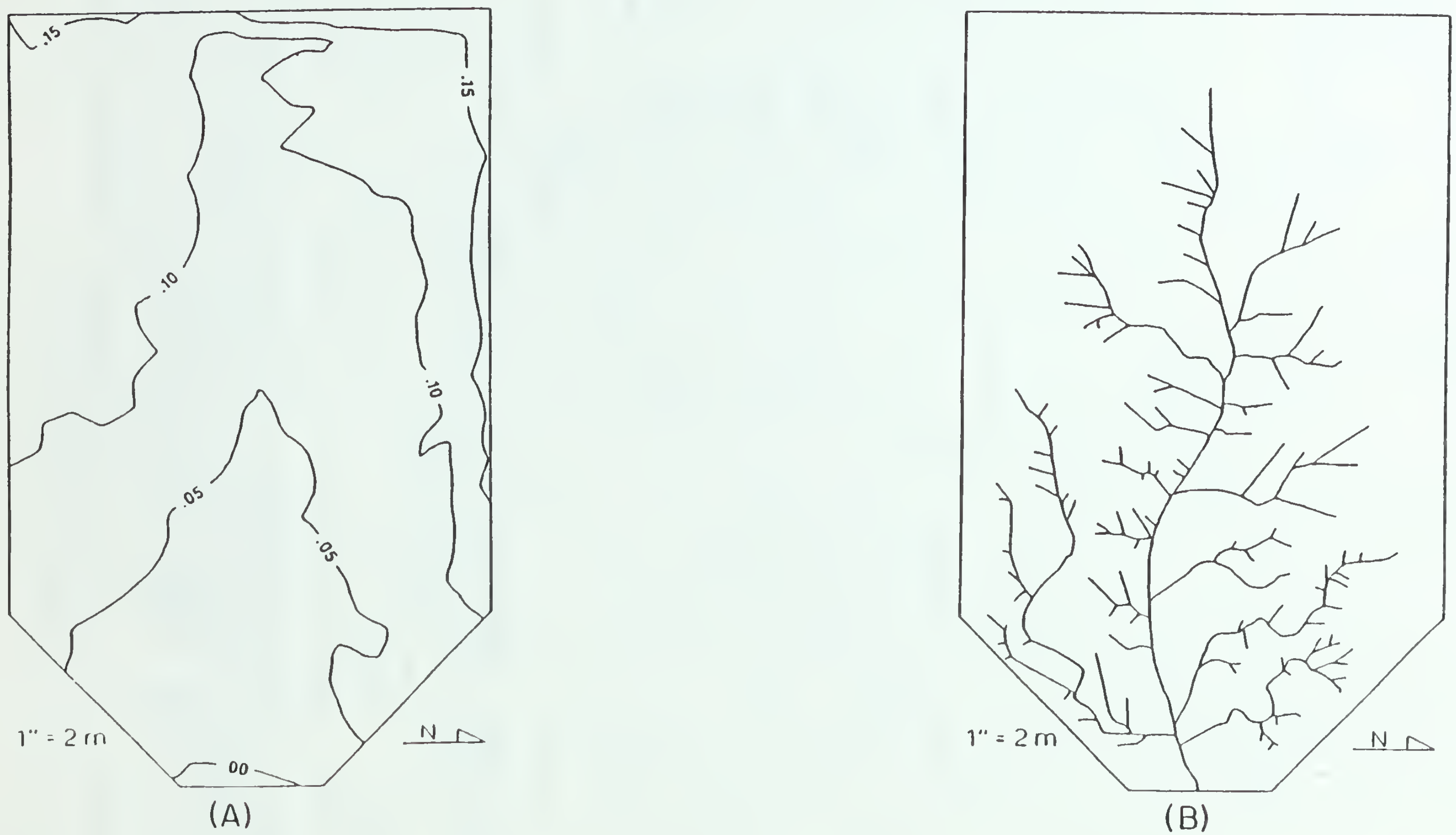

Figure 3-4. Contour map of initial basin surface (a)(Contour Interval - 0.05m) and experimental channel network after 1022 mm cumulative precipitation. (McLane, 1978)

The conclusion to be drawn is that if a map of an undissected surface has been prepared in some detail then it should be possible to determine where channels will form. In fact, if one compares the network that developed during this experiment with the original contour map (Fig. 3-4) it appears that even the major tributaries are influenced by irregularities of the original surface. This is important because it indicates that it is possible to identify those sites where channel extension or gullying will occur by careful mapping of an undissected surface.

Figure. 3-5 shows the growth rate of the network, as reflected by total channel length. During phase 1 of the experiment, prior to the lowering of the base level, the decrease in the rate of network growth is apparent. This is the result of declining rates of runoff at the head of the advancing channels. The farther the nickpoint advances from the outlet of the REF the less water moves over the nickpoint. In other words, the head of the channel is running out of contributing areas and, therefore, its rate of extension will slow.

The network growth trends are reflected in sediment yield (Fig. 3-6). The initial erosion occurs at the outlet of the container, and the growth of the drainage network is headward. Therefore, almost all of the sediment from initial channel incision is exported from the REF, and sediment concentrations are very high. However, as the network grows the rate of growth decreases, and the zone of maximum sediment production moves headward with the nickpoints. The opportunity for sediment storage in the downstream channels is increased and sediment concentration decreases markedly with time to an average value reflecting basin morphology and sediment erodibility.

BASE-LEVEL LOWERING

Phase 2 of the McLane study involved rejuvenation of the drainage system by a 5 cm lowering of base level. Base level was not lowered until the network had developed to a maximum for the given slope and precipitation characteristics. Immediately following base-level reduction, a large nickpoint formed at the outlet, and it moved up the main

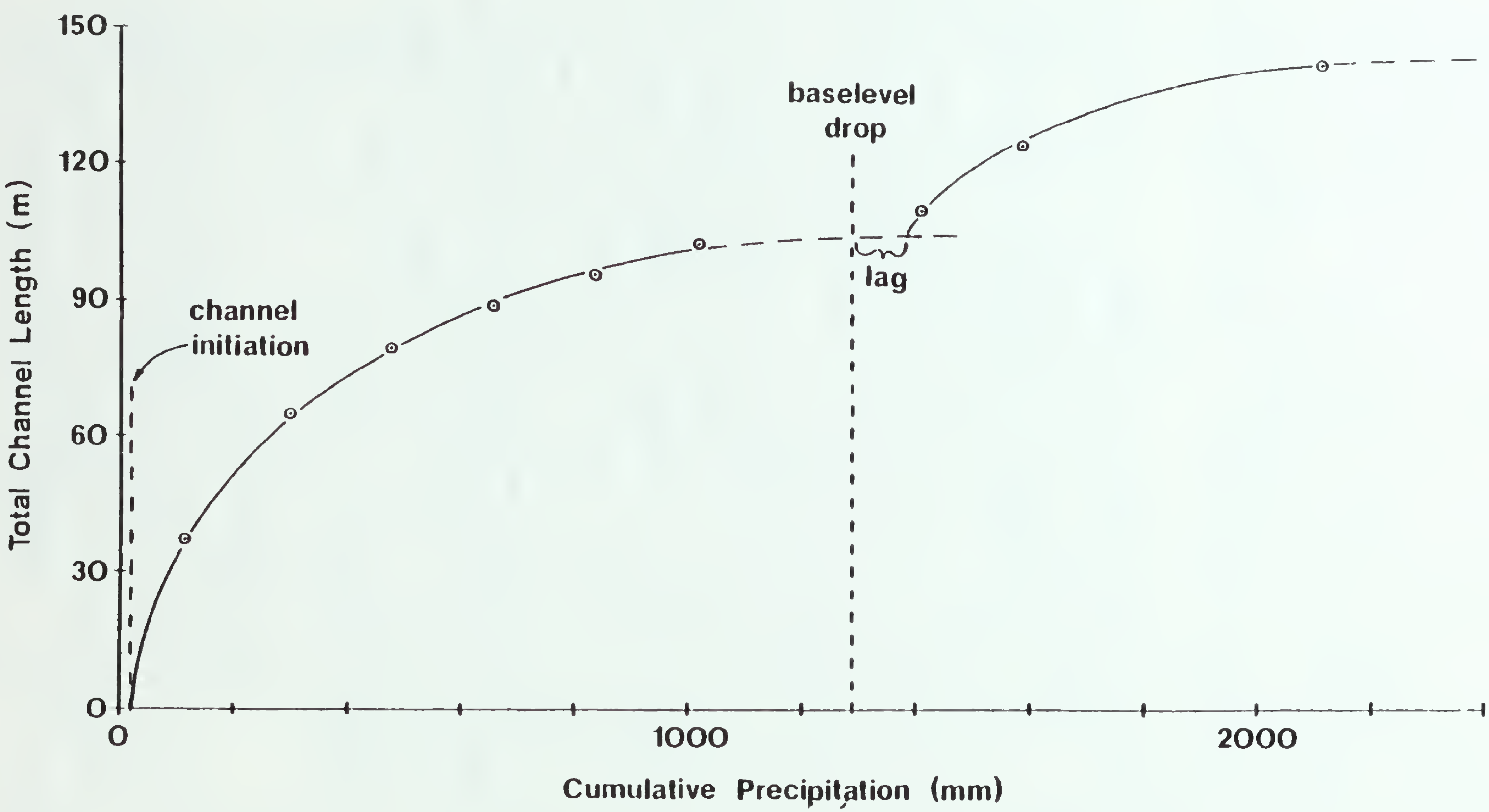

Figure 3-5. Plot of total channel length and cumulative precipitation vs. a time index. Rapid initial increase followed by gradual slowing is evident during both growth phases. (McLane, 1978)

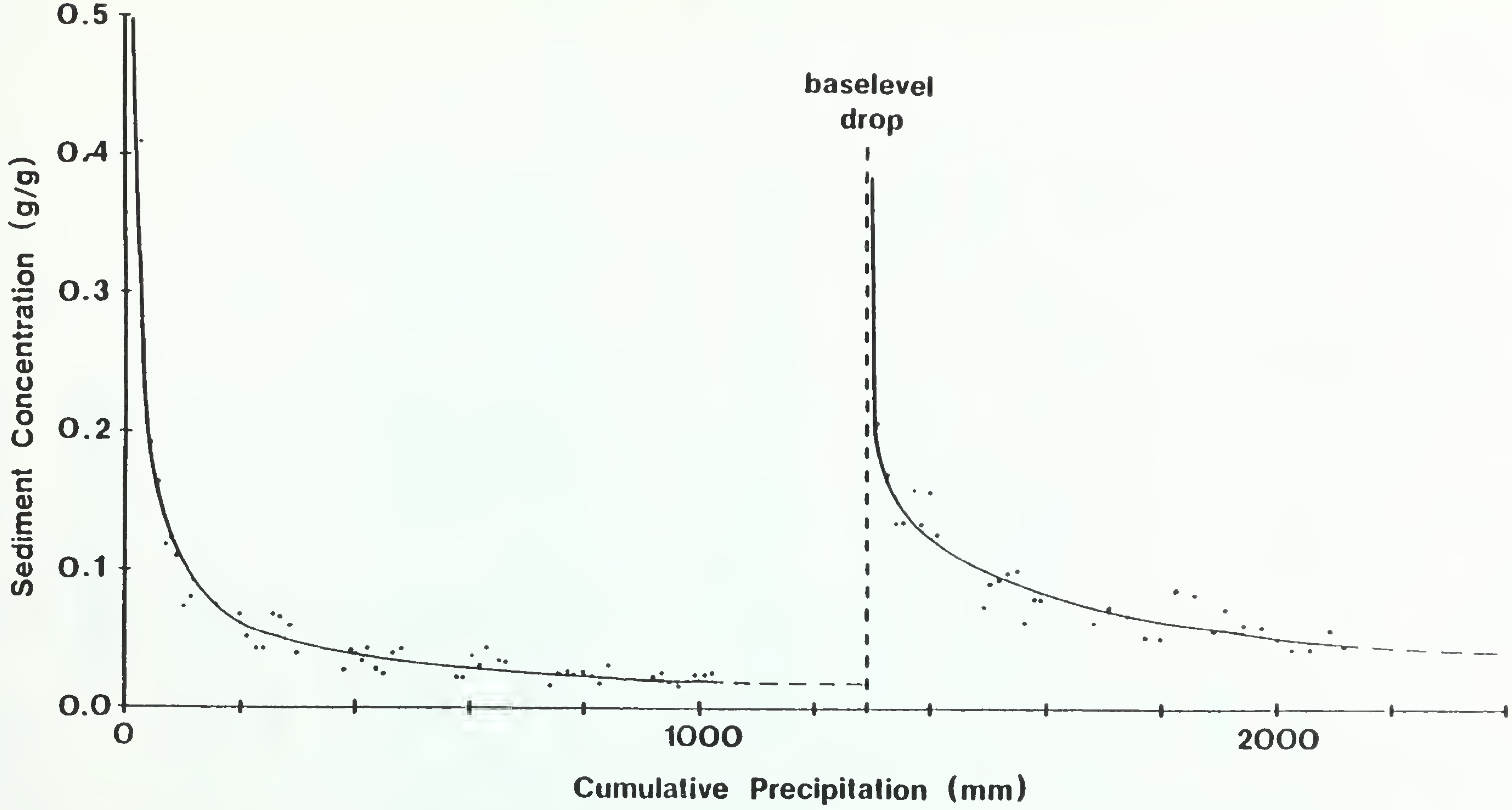

Figure 3-6. Plot of sediment concentration and cumulative precipitation. Exponential decrease during both growth phases reflects decrease in erosive power. (McLane, 1978)

channel. Although sediment yields immediately increased (Fig. 3-6), the drainage pattern was not affected until the effect of the rejuvenation was felt at the channel heads. A lag in network response is shown on the right side of Figure 3-5. This indicates that there can be a very significant lag time in a natural system between the downstream effect and its upstream response. Therefore, in some cases years may pass before the activities of man in the downstream part of the basin will affect the upstream reaches.

In summary, the McLane (1978) study shows the effect of base-level lowering on drainage network growth and on sediment production. It also indicates how important is the initial configuration of the surface, and McLane also concludes that the growth of the drainage network is controlled by the critical contributing area, which is established in the basis of eroding forces and the resistance of the material to channelized erosion. This concept of a critical contributing area will be considered in more detail later.

McLane's (1978) experiments were terminated, when the drainage network had achieved maximum extension, but an earlier experimental study in the same facility (Parker, 1977) was carried beyond maximum extension and a significant decrease of drainage density was recorded as first-order channels were eliminated (Fig. 3-7). As in McLane's study an increase of relief and slope increased sediment yield. The variability of sediment yield was high as Figure 3-8 demonstrates. The rapid decrease of sediment yield following base-level lowering is characteristic, but, in addition, secondary peaks of sediment yield occurred as the network evolved. This has been referred to earlier as complex response, and it is related to secondary nickpoints (Fig. 1-1) or channel scour removing stored sediments. Shorter periods of high sediment production are related to periods of bank collapse and to periods of rapid primary-nickpoint migration.

NICKPOINT MIGRATION

In the experimental facility a drop of base level produced nickpoints, which migrated up the main channel and its major tributaries

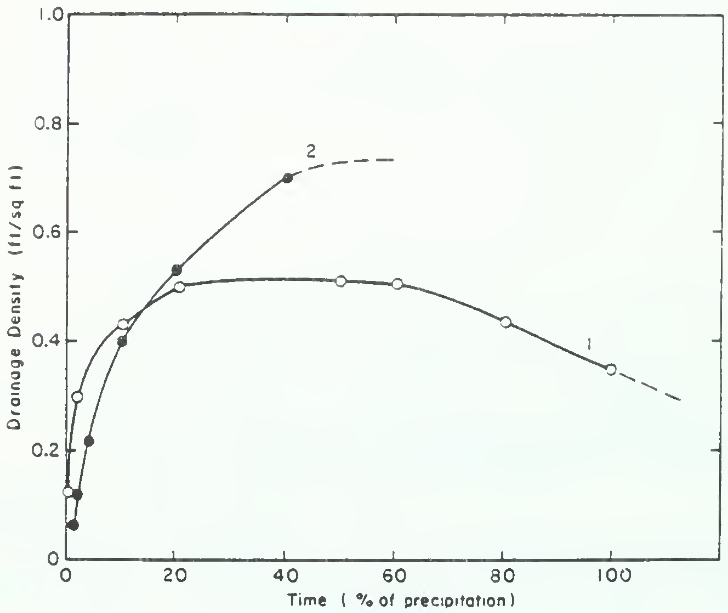

Figure 3-7. Drainage density change during two experiments in REF. Curve 1 is for pattern developed on 3.2 percent slope with stable base level. Curve 2 is for pattern developed on 0.75 percent slope with base-level lowering prior to beginning of experiment. (After Parker, 1977). Time is expressed as a percentage of the total water applied to the REF during experiment 1 (curve 1).

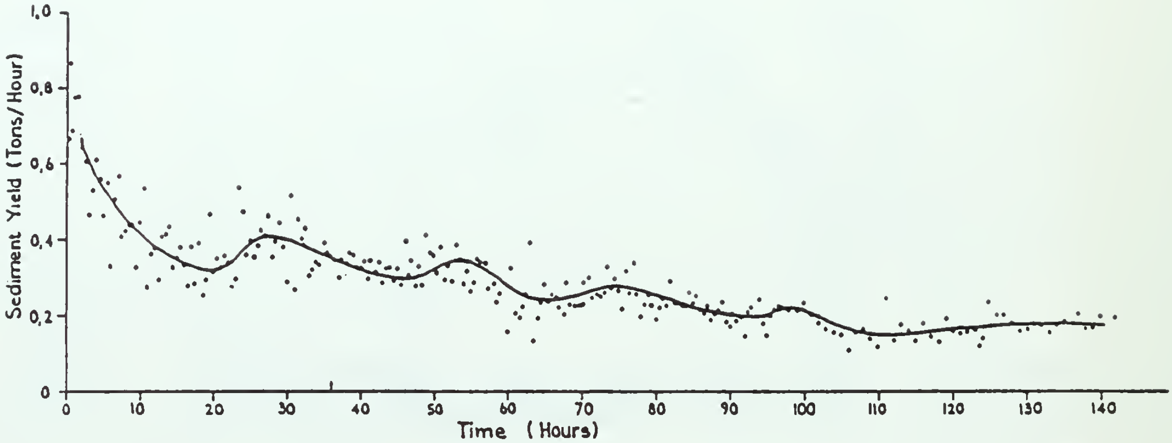

Figure 3-8. Moving average sediment yield for experiment with an initial slope of 3.2%. (Parker, 1977).

(Parker, 1979). As a nickpoint migrated upstream, the drainage area above a nickpoint was not influenced by the change in base level, whereas downstream erosion was accelerated in the channel and along the valley walls in response to the base-level changes. Nickpoints, therefore, represent the position of the "wave of dissection" produced in response to the base-level change.

Nickpoint migration in the main channel was measured, and it shows a decreasing rate of migration with time or distance up the main channel (Fig. 3-9). The equation is:

$$K = 34.48\ V^{-.99} \tag{3-1}$$

where K = nickpoint migration in ft/cu ft of water delivered

V = volume of water delivered to the system (ft^3)

Figure 3-10 records nickpoint movement up the main channel and three tributaries. The upper curve (Fig. 3-10) shows nickpoint migration along the main or central stream in the watershed. The series of lines branching from this and located below it represents nickpoint migration in three tributaries. Nickpoints slowed at lower order tributary junctions, although migration along the higher-order channel continued.

Seginer (1966) in a review of gully-erosion research found that size of the drainage basin figured prominently in predicting gully advance. He suggested that gully advance could be evaluated from an equation of the form:

$$R = cA^b$$

where R = average annual linear gully advance

A = area of drainage basin

c = constant

b = constant

When the data from the REF experiment are plotted the result is an exponential relationship (Fig. 3-11).

The independent variables of watershed area, distance from the basin outlet and stream order are all highly intercorrelated, and of course, they are related to discharge. As a nickpoint moves into a watershed, its rate of migration is slowed because of the reduction in basin area

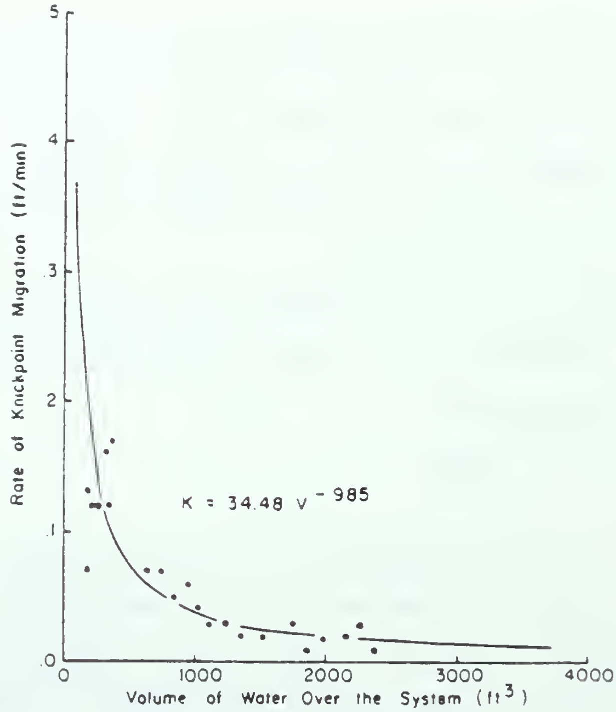

Figure 3-9. Rate of nickpoint migration through time. (Parker, 1979). Time is expressed as volume of water applied.

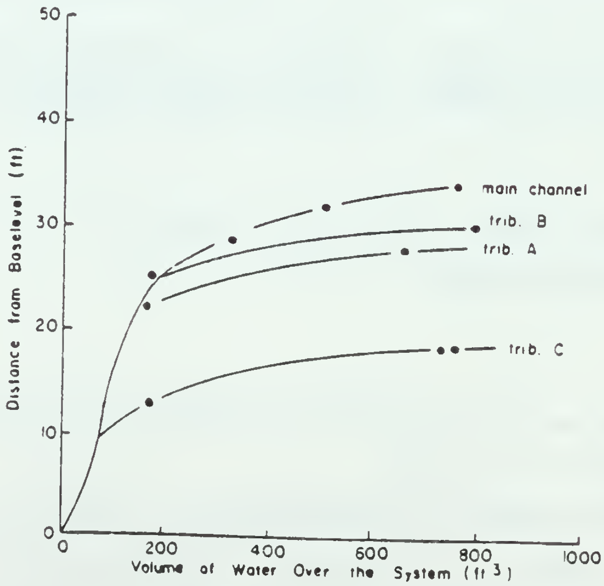

Figure 3-10. Nickpoint migration shown as distance from the basin outlet after a base-level change. Time is expressed by volume of water over the system. (Parker, 1979)

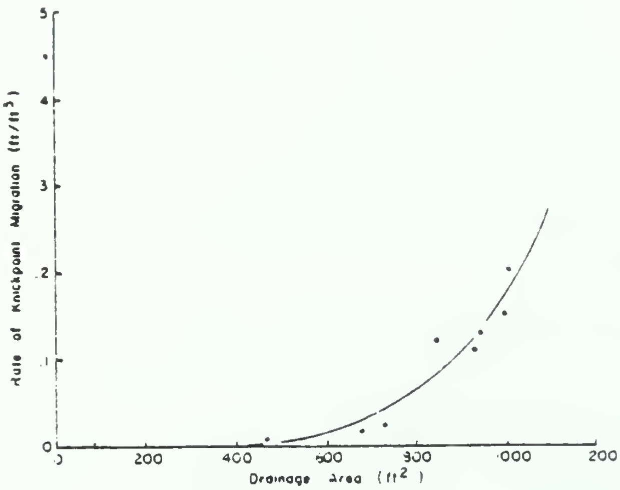

Figure 3-11. The relation between the rate of nickpoint migration and the approximate drainage area above the nickpoint. Because volume of water is used as an index of time, the rate of nickpoint migration is shown in feet per cubic foot of water. (Parker, 1977)

and the resulting decline of discharge, which are more significant than the headward increase in channel slope (i.e. concave profile).

4. GULLIES

Gullies have been defined earlier in Chapter 2 as incised channels that form where no well-defined channel previously existed. Brice (1966) in an excellent study of the morphology and gully characteristics of the Medicine Creek drainage basin, southwestern Nebraska, classified gullies into valley-floor, valley-side and valley-head gullies. Each type can be discontinuous or continuous (Figs. 4-1, 4-2). Continuous gullies form part of a drainage network, whereas discontinuous gullies are isolated (Fig. 4-1). Valley-side and valley-head gullies are basically the same, and they reflect an expansion of the drainage network. Valley-floor gullies reestablish a drainage channel in the alluvial floor.

VALLEY-SIDE GULLIES

A classic report on gully development and evolution is that of Ireland et al (1939), which describes the very serious gully problems in the Piedmont of South Carolina during the first third of this century. In high precipitation areas dense vegetation should prevent serious erosion, however, it is in these areas that erosion will be greatest when the natural vegetation is removed. In addition, the torrential nature of hurricane precipitation in the southeast is another factor that greatly aggravates the erosional problem.

Almost all gullies result from increased runoff or from an unnatural concentration and acceleration of flowing water (Table 2-1). In many cases, the shape of the gully that forms is related to its cause of formation (Fig. 4-3). In contrast to the development of gullies in relatively homogeneous loess, the formation of gullies in the Piedmont is significantly affected by the soil horizons. The B horizon is resistant and acts as a caprock, whereas the underlying C horizon is composed of weak-weathered rock. The end result is that a breaching of the B horizon leads to rapid gully enlargement.

Ireland et al (1939, p. 137) identify four stages in the evolutionary development of the gullies as follows:

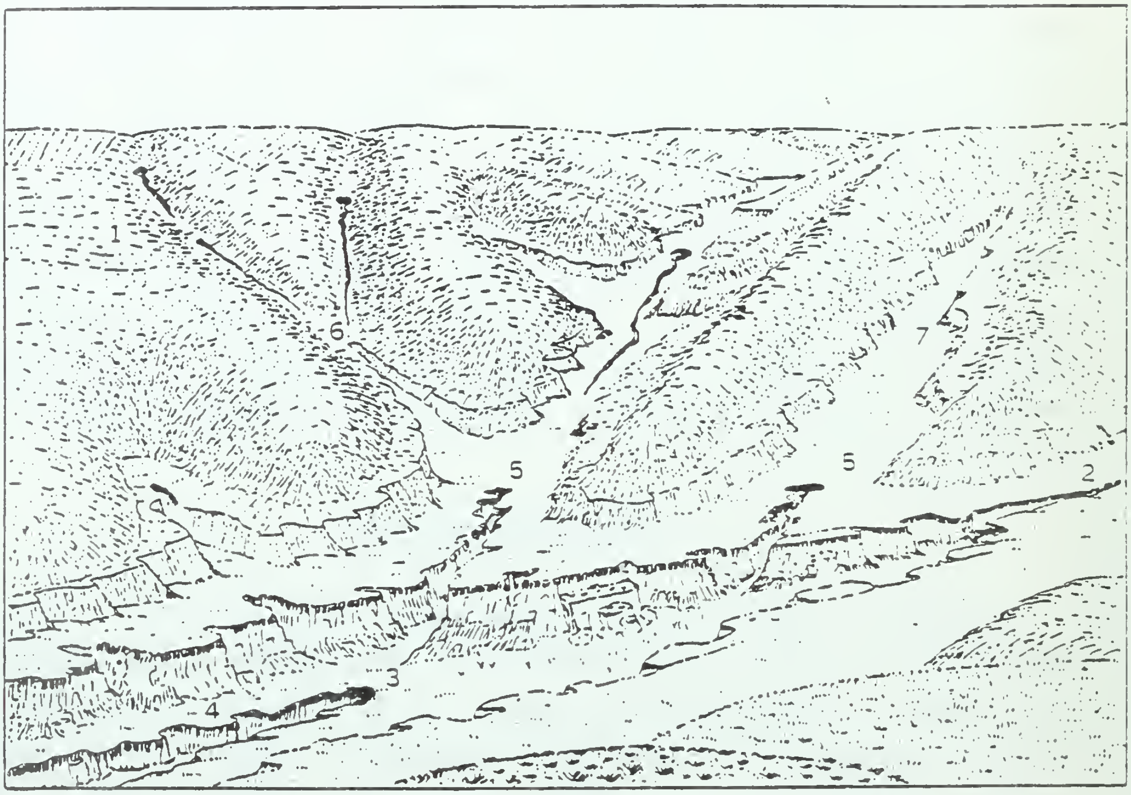

Figure 4-1. Composite sketch, based on field sketches and photographs, showing the different varieties of valley-floor gullies. Actively eroding scarps are indicated by darker shading. Discontinuous gullies (1) form in tributary valleys. They will coalesce when the headcut at 6 reaches them. In this valley there are two nickpoints (2, 3). The incision of the valley floor forms terraces (4, 5). Tributary gullies erode alluvium that fills tributary valleys (7) and they will eventually integrate the drainage network (Brice, 1966).

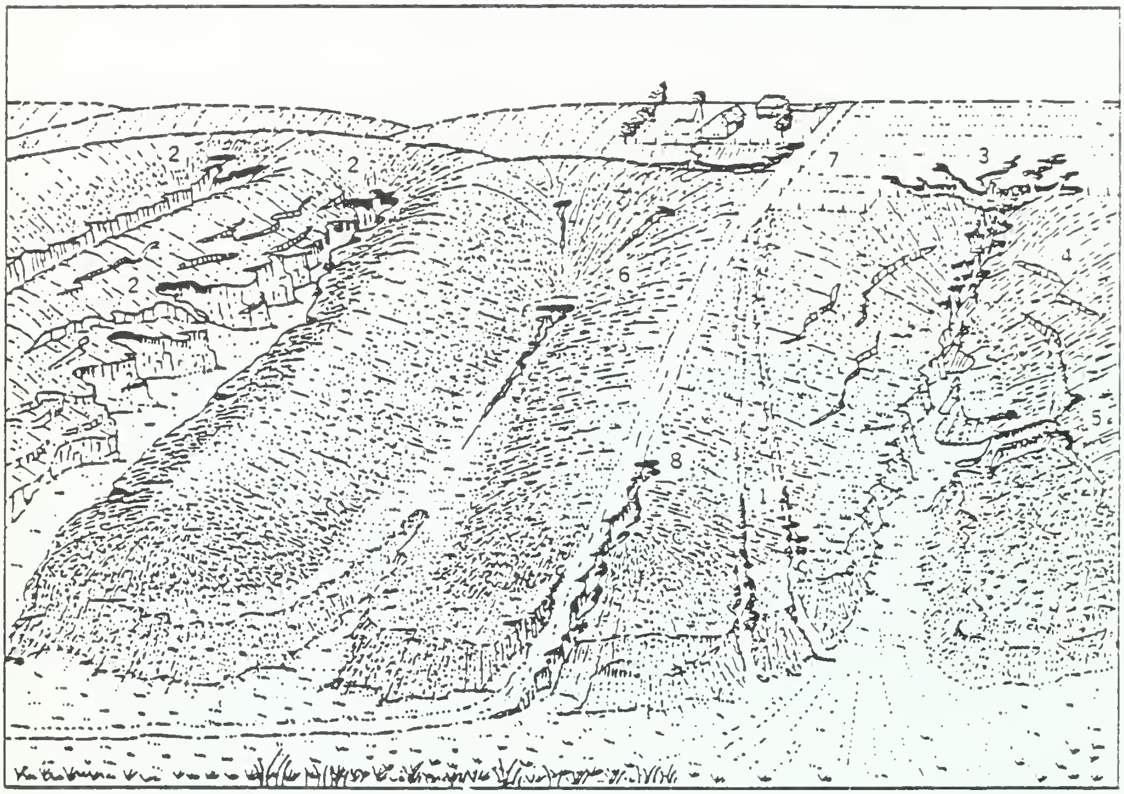

Figure 4-2. Composite sketch, based on field sketches and photographs, showing different varieties of valley-head and valley-side gullies. Actively eroding scarps are indicated by shading. Discontinuous gullies are again present (1, 6) are not in alluvium. Valley-head gullies are either lobate (2, 6) or branching (3) or sharply defined (1, 3) depending on concentration of flow. Valley-side gullies could also have similar morphology but Brice shows them as branching (6) or straight with a well-defined headcut because they follow roads (7) and trails (8)(Brice, 1966).

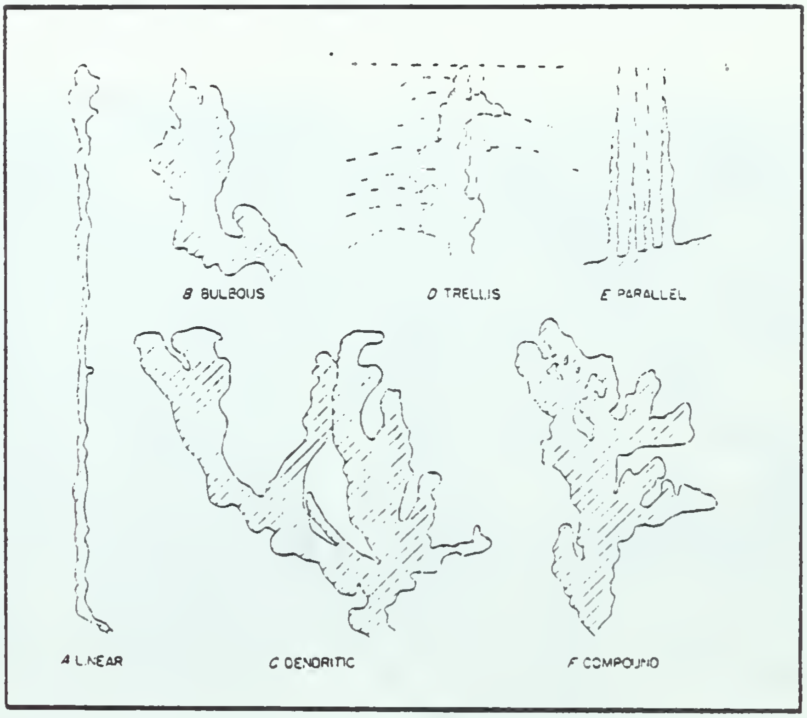

Figure 4-3. Character of gully forms:

A. Linear: long and narrow, with narrow head and few important tributaries along its sides; common along property lines; follows old or existing drainage ditches, many of which have been dug or plowed out to serve as terrace outlets. Linear gullies may grow broader at the head and with further growth may develop a bulbous, dendritic, or compound pattern.

B. Bulbous: Broad and spatulate at upper end, but may be linear in downstream portion; incised in upland, often following the course of an old natural drainage having a semicircular or amphitheatre-shaped head with small tributaries or rills entering from all sides. The pattern is likely to become dendritic as the gully grows older.

C. Dendritic: Formed of many branching tributaries; usually developed following the natural drainage lines but may be due in part to ditches, terraces, road drains, in a semicircular or amphitheatre-shaped head. Headward cutting along tributaries accentuates the dendritic character.

D. Trellis: Tributary gullies or branches enter the main channel at angles approaching 90^{o}; developed on a flat or evenly sloping area where a system of terraces empties into a central terrace drain or outlet. Headward erosion along the terraces further accentuates the trellis effect.

E. Parallel: Composed of two or more parallel tributaries which empty into a main gully, as with drainage of old roads where ruts and road ditches run parallel for some distance before coming together into a main gully channel. Headward cutting accentuates the parallel development for a time but capture of one tributary by another is likely eventually to destroy the parallel pattern.

F. Compound: Combinations of any two or more of the gully forms. The main channel may have tributaries entering in trellis pattern and tributaries themselves may be dendritic, parallel or bulbous. (Ireland, et al., 1939)

"In stage 1, the channel-cutting stage, the gully works downward through the A and B horizons. Cutting in this stage is relatively slow, and this is the time at which protective measures can best be undertaken.

Stage 2 begins when the gully penetrates the base of the B horizon and begins cutting in the weak parent material. This stage, characterized by the headward migration of an overfall and plunge pool and by rapid caving of the walls and deepening of the channel, is much the most violent stage of gully growth and is the least favorable for the successful application of control measures. Caving and slumping of the walls and heads alternate with periodic clearing out of the caved material from the gully channel. Additional substages in the gully's growth may occur in the form of periods of headward progression of successive waterfalls or knickpoints, marking renewed channel cutting and deepening of the gully. Stage 2 ends when erosion is retarded because the channel reaches a graded condition under the control of some local base level.

Stage 3 is a period of adjustment to the graded channel. Slopes of the gully walls are reduced by weathering, slope-wash and mass movement; plants are able to get a foothold on the lowered slopes, and vegetation gradually brings about a healing of the gully.

Stage 4 is a period of stabilization and is characterized by the slow development and accumulation of new top-soil over the old scarred surface. Rejuvenated cutting brought about by lowering of the base level or an increase in the amount or rate of runoff can at any time cause stage 3 or 4 to revert to stage 2."

The four stages of the gully cycle pertain not only to one reach of a gully through time but also the differences in gully morphology from mouth to head. Hence, if a series of cross sections are surveyed they can be arranged from mouth to head to show the evolutionary sequence of gully change with time.

In spite of the great differences in climate and the nature of the materials, the evolution of the Piedmont gullies seems to be very similar to that described by Bariss (1971, 1977) for gullies in the loess of central Nebraska. Each of the stages of gully development is characterized by a typical cross section. In the first or initial phase there is no continuous channel, but small discontinuous gullies or collapse holes related to piping may be present.

According to Bariss (1977) the initial phase is followed by an "unstable channel phase" during which a continuous channel incises to a local base level. The channel floor is irregular and steep unstable side-walls have slopes in excess of 45^{o}. This is a phase of active gully development like the Ireland et al (1939) stage 2 and development is very rapid.

A "phase of stabilization" begins after channel incision ceases. The gully widens and the side-slopes decline to about 40^{o} (Fig. 4-4). The maximum side-slope angles range from 26 to 42 degrees, with a mean of 36^{o}.

From the investigations of Ireland et al (1939) and Barris (1977), it is clear that valley-side gullies evolve to a condition of relative stability when the channel has cut to a new base level and the gully walls have reclined and vegetation has colonized the gully sides.

In the gullies studied in the Spartanburg area a wide range of gradients were found. The gradient was dependent on many factors, such as the position in the gully, character of the soil material and degree of activity in the gully. For example, a tributary may introduce large quantities of sediment to the main gully channel, which is stored and forms a valley fan or plug convexity on the long profile. The longitudinal profiles and cross sections presented by Ireland et al (1939) show very clearly how complicated is the erosional evolution of these gullies. Two of their gullies will be considered here, Layton's Gully and Waldon's Gully. Layton's gully is not as old as many of the other gullies in the area, but it has passed through several distinct stages in its development. It shows a marked contrast between portions which are now temporarily stabilized and others which are cutting with great rapidity.

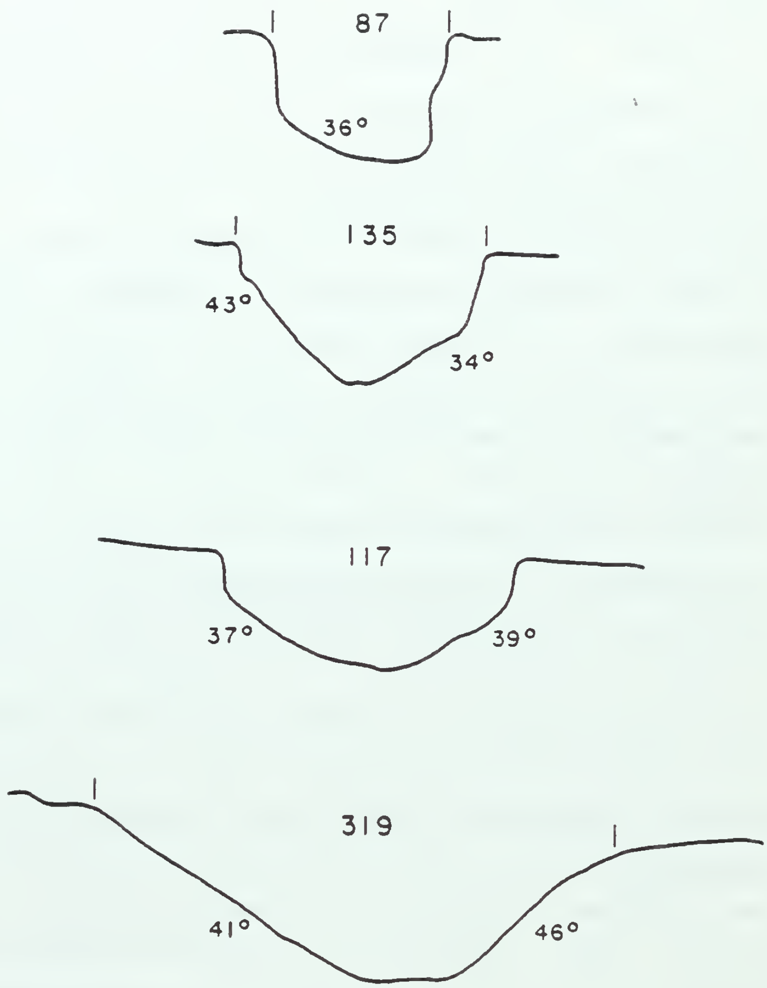

Figure 4-4. Gully cross section evolution in central Nebraska. Number above cross section is width of gully in feet and other numbers are slope in degrees. (Bariss, 1972)

For convenience of discussion Ireland et al (1939) divide Layton's Gully into four sections, each having different erosional characteristics (Figs. 4-5a). The following is a summary of their discussion (Ireland et al, 1939, p. 90-104).

The upper or headwater segment is a little more than 90 feet long and extends about to the position of cross section 3 (Fig. 4-5a). This portion of the gully is less than 23 years old, and it occupies an area that was formerly terraced and cultivated. It is now almost barren of vegetation and is actively eroding. The walls are retreating by washing, crumbling, slumping and caving. Individual cutting heads are progressing by plunge-pool action, by washing, and where weak C material is softened by back trickling of sheets of water along the vertical or overhanging walls, by caving. Caved material is removed from the gully channel only when heavy rains cause abundant runoff.

The middle segment extends from cross section 3 to section 9 and comprises some 350 feet of fairly well stabilized gully much older than the upper segment. In this stretch, the stabilized character of the floor and walls and the age of the pines and tulip-trees now growing there indicate that the channel has existed for at least 35 years. Portions of this segment contain fresh scarps and small terraces that indicate active erosion.

The stabilized condition of the middle segment of the gully will soon be disturbed. The lower part of this segment contains numerous small secondary nickpoints that developed as a result of rejuvenation of the lowest segment of the gully. The youngest and largest of these was formed when runoff from the storm of October 15-16, 1936, caused erosion of the main gully. At the start this nickpoint was 2 feet high but by August 13, 1937, it had migrated upstream 42 feet and had increased to a height of more than 4 feet. Ireland et al (1939), conclude that headward movement of this nickpoint is certain to continue unless protective measures are taken or unless it encounters a resistant dike or other rock barrier. The nickpoint will destabilize the middle segment.

The long profile (Fig. 4-5a) is very revealing. Note that the lower end of the natural valley floor below 12 was convex indicating a

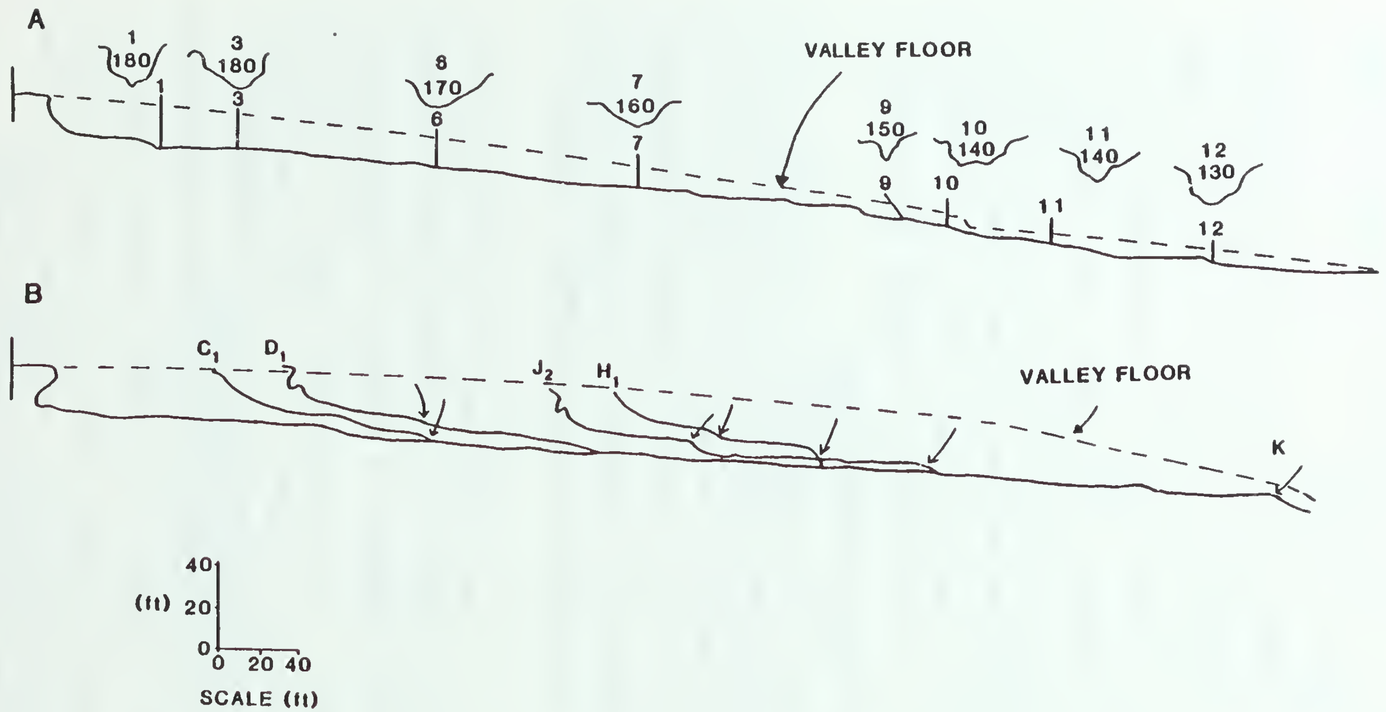

Figure 4.5. Gullies of South Carolina piedmont (Ireland, et al, 1939)

a. Layton's Gully, longitudinal profile and cross sections. Numbers above cross section are section number and top width in feet.

b. Walden's Gully, longitudinal profile of main channel and important tributaries. Open arrows show positions of nickpoints.

condition of relative instability. Note also the numerous nickpoints and convexities in the profile of the main gully channel. These are evidence of continued instability, with erosion and deposition alternating within the channels at any given reach.

Walden's Gully which is one of the most spectacular in the area (Fig. 4-5b). Its deep canyon-like channels are incised as much as 35 feet below the evenly sloping upland, portions of which have been completely isolated and left as islands surrounded by steep, vertical or overhanging gully walls.

Walden's Gully appears to have started about the middle of the last century. By 1855 it was about 20 feet deep. It is of compound form. Headcuts working up the slope have tended to produce a dendritic or branching pattern. Erosion cutting headward along terrace furrows, since 1900 or 1905, has carved lateral tributaries to the gully, thus giving part of it a trellis pattern (Fig. 4-6).

The channels of several of the basins contain nickpoints, most of which have developed in the main channel and have worked headward up the tributaries. Some of these are 2 to 5 feet high and make the channel a series of steps (Fig. 4-5b). Many of them may be correlated with individual storms or times when cultivation was intensified in response to high prices or increased demand.

An important conclusion to be reached concerning the studies of the South Carolina and Nebraska gullies is that they do progress through identifiable stages from inception to stability, but the evolution is not progressive. It will be interrupted by periods of deposition and erosion. In effect the evolution is episodic. The irregularities in the long profile of Layton's and Walton's Gullies are a clear indication of their complex evolution.

VALLEY-FLOOR GULLIES

Valley-floor gullies become valley-side gullies as they advance headward into previously ungullied areas. The valley-floor gullies, that are incised into alluvium, become conduits for the downstream movement of water and sediment. They develop by base-level lowering or

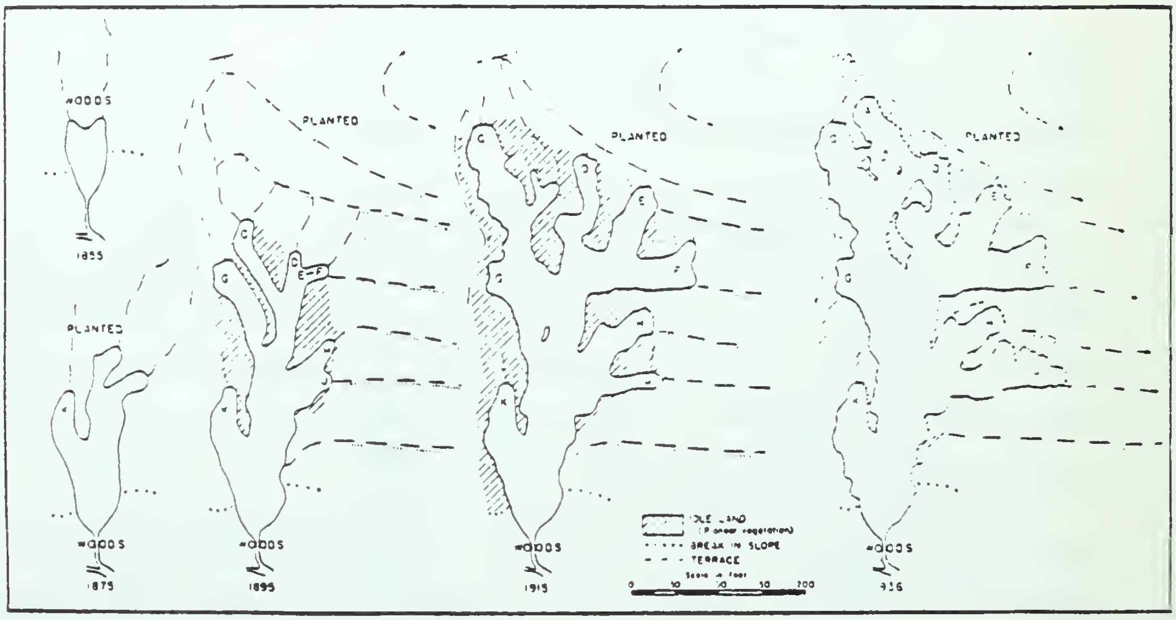

Figure 4-6. Reconstructed history of Walden's Gully, 1 miles west of Moore, S.C. (Ireland, et al., 1939).

by discontinuous gully formation. Discontinuous gullies, as described earlier, are frequently an indication of valley-floor instability and a harbinger of further erosion. For example, Everitt (1979) in his study of the Bull Creek arroyo in Utah found that he could, by the use of tree ring dating techniques, date the origin of various reaches of Bull Creek. He concluded that the midsection was the most recently formed, indicating that there were discontinuous gullies in the upper and lower reaches of Bull Creek, which were integrated by incision between them.

Leopold and Miller (1956) discuss the formation of discontinuous gullies in the smooth valley floors of New Mexico. They indicated that any local weakening of the vegetation permitted the formation of a discontinuous gully and because the channel was not transporting large quantities of sediment, it cut back at a much gentler gradient than the valley floor itself. They further indicated that discontinuous gullies coalesce to form a continuous gully system of steeper gradient and, of course, larger dimensions. Brice (1966) provides an example of the coalescence of two discontinuous gullies in the Medicine Creek area (Fig. 4-7). This coalescence took place by the rapid advance of the downstream headcut into the upstream discontinuous gully, thereby integrating the two.

Heede (1967, 1974) makes the point that discontinuous gullies are representative of the most youthful stage of valley instability and gully development. At least in the western United States, he considers discontinuous and fused gullies as representative of youthful and early-mature stages of gully development.

DISCONTINUOUS GULLIES

Discontinuous gullies are particularly easy to study because they are isolated from the rest of the drainage network. This makes it possible to relate the discontinuous gully to the characteristics of the valley floor. For example, Schumm and Hadley (1957) identified steep critical locations on a valley floor, where the initial incision occurred. They demonstrated that alluvium accumulates in semi-arid valleys until a threshold slope is exceeded, and a discontinuous gully forms on the steeper reach of the valley floor.

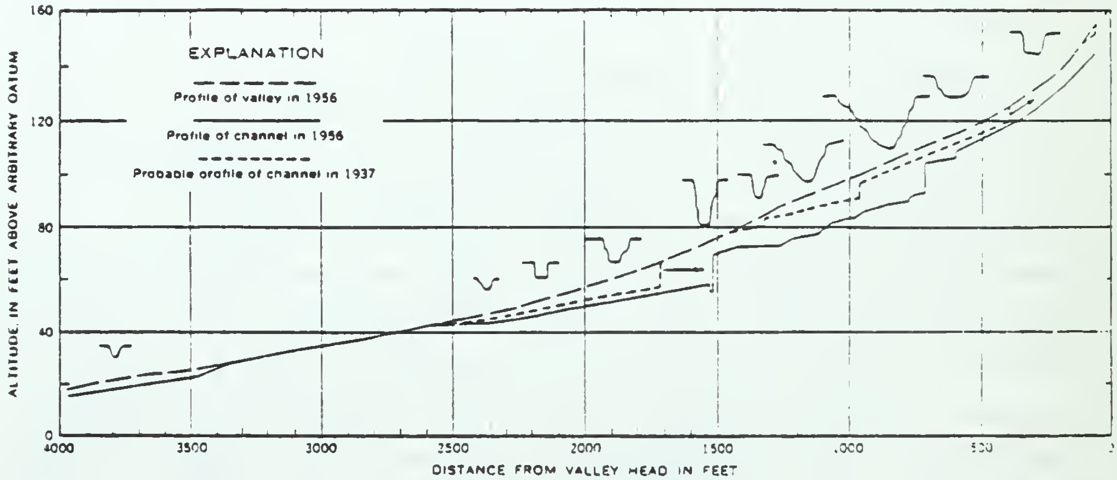

Figure 4-7. Changes in long profile of a gullied tributary to Dry Creek, 1937-56 (Brice, 1966).

Observations in Wyoming and elsewhere reveal that one other important characteristic of a semi-arid drainage system is the lack of accordance of tributary and main channel, that is, the tributaries may or may not be graded to the main channel. An example of this is the Twenty Mile Creek drainage basin where one-third of the tributaries are graded to the surface of a terrace above the main channel (Hadley, 1960). This lack of integration in certain channels is due to the filling of the lower parts of the tributary valleys with sediment, thereby causing spreading of water and sediment over the terraces and floodplain adjacent to the main channels. In any larger drainage system the tributaries may be found in all stages of integration with the main channel suggesting that each tributary has its own history of alluviation and dissection which may not be contemporaneous with that of its neighbors.

Field examination of a number of tributaries of Twenty-Mile Creek show in general two cases: (1) actively eroding headwater channels and filled or alluviated lower valley reaches and (2) moderate erosion in headwater channels and gullies trenching the lower valley fill. These two seemingly distinct examples appear to be the two components of the semi-arid cycle of valley development. Schumm and Hadley (1957) used these observations to develop a model of semi-arid erosion as follows: "An alluviated tributary valley is united with the major drainage channel by the development of a trench in the recent alluvium clogging the tributary channel. The gully is extended headward by upstream headcut migration. Figure 4-8A shows the channel at the beginning of this rejuvenation. As the headcut migrates up channel the lower section of the tributary drainage (Sec. 1, Fig. 4-8A) becomes very efficient for sediment transport, for the runoff is concentrated in a clearly defined channel. The headcut continues to work up channel, passing tributaries 2 and 4 and rejuvenating them in turn (Fig. 4-8B). Runoff increases and time of concentration is much shorter, but the volume of sediment moved is greatly increased by the rejuvenation of tributaries 2 and 4. With the rejuvenation of tributaries the upland slopes may be steepened and they in turn may supply more sediment to the channels. Thus, as the

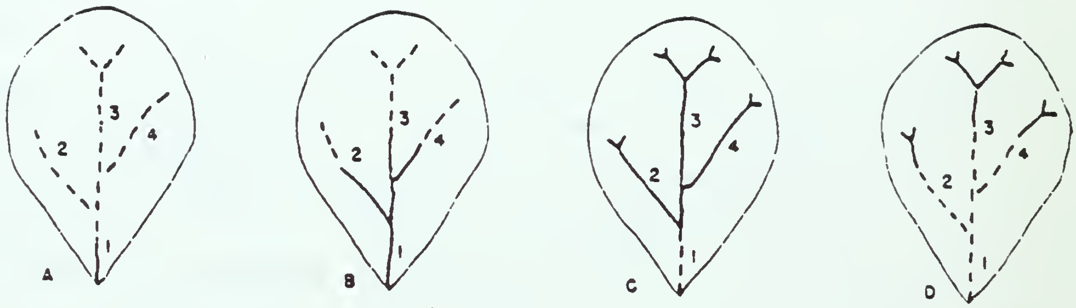

Figure 4-8. The cycle of trenching and alluviation in a semiarid valley. Dotted line indicates alluviation within a channel. Solid line indicates trenching or a well-defined channel (Schumm and Hadley, 1958).

headward cutting continues, upchannel sediment delivered to section 1 of this basin is greatly increased as is the velocity of flow but with perhaps only minor increase in runoff. Sections 2, 3, and 4 now are supplying much sediment, but section 1 has a gradient formed when the supply of sediment was low. Aggradation begins in section 1, perhaps due also to widening of the channel (Leopold and Miller, 1956), or deposition may begin at the junction of tributary with the master stream (Fig. 4-8C). Thus, aggradation begins as discontinuous deposits in reaches of the valley with either the least slope or greatest width or both. Alluviation once begun promotes more alluviation until the channel is filled and is no longer smoothly graded to the main drainage channel. The area of maximum channel deposition then migrates upstream. As it passes the mouths of tributaries they begin filling their valleys, for the local base level is raised. The area supplying sediment to the main channel decreases in size (Fig. 4-8D) and the rate of alluviation undoubtedly decreases. The gradient of the filling valley becomes steeper until discontinuous gullies begin to form in stretches of channel where the gradient is steepest. These discontinuous gullies shift the alluvium down channel by building fans in the valley which are in turn trenched when the gradient steepens. Finally the downstream movement of sediment forms a fan near the tributary mouth. This fan is in turn trenched, integrating the tributary drainage with that of the master stream, and so the cycle begins again."

Blong (1966, 1970) in New Zealand, Tuckfield (1964) in England and Brice (1966) and Thornthwaite et al (1942) recognize that discon-tinuous gullies form on steeper critical reaches of the valley floor. These investigators have documented changes in gully cross section through time which conform to observations made by Ireland et al (1939). All of this work emphasizes the considerable complexity of gully evolution.

Piceance Creek Area

If discontinuous gullies form on oversteepened convexities on the valley floor these critical locations can be identified, and preventive measures may be undertaken to inhibit the formation of the gully at that location.

An effort was made as part of an investigation of the environmental effects of oil-shale mining (Patton, 1973) to identify these critical locations in valleys of the Piceance Creek basin of western Colorado, and the previously discussed Figure 2-1 shows the result. Patton and Schumm (1975) used two simple geomorphic variables drainage area and valley slope, to define a discriminant function distinguishing between gullied and ungullied valleys.

In the absence of hydrologic data drainage-basin area was used as an index of discharge (Burkham, 1966). The data were plotted on semi-logarithmic paper on which the discriminant function is a straight line (Fig. 2-1). The line represents a relationship of the type:

$$K = aAe^{-s}$$

where A is a drainage area, s is valley slope, e is the base of natural logarithms, and a is a constant. Here K should be a certain threshold parameter, its value being constant along the discriminant line. However, such a parameter does not bear a clear relation to hydraulic variables such as shear stress or stream power, and the distance of a point from that line is but a qualitative measure of valley instability.

In these valleys the flat-appearing alluvial valley floors are irregular in a downstream direction, that is, there are convex reaches where sediment is stored (Fig. 4-9). These reaches frequently occur at or downstream from tributary junctions, and within one valley there may be numerous such convexities. The downstream nose of the convexity, which may resemble an in-valley alluvial fan, is the steepest reach of the valley floor, and this is the gradient that was measured. The drainage area above this reach was measured, and these two values provide one data point on Figure 2-1.

Semi-arid valleys of one area frequently are in various stages of stability, gullying and healing and by a study of each of these valley reaches a quantitative assessment of valley stability can be obtained.

The following is a reanalysis of the work of Patton and Schumm (1975) by Begin and Schumm (1979). Patton (1973) measured valley slope and drainage area in 56 valleys. In 23 valleys either continuous or

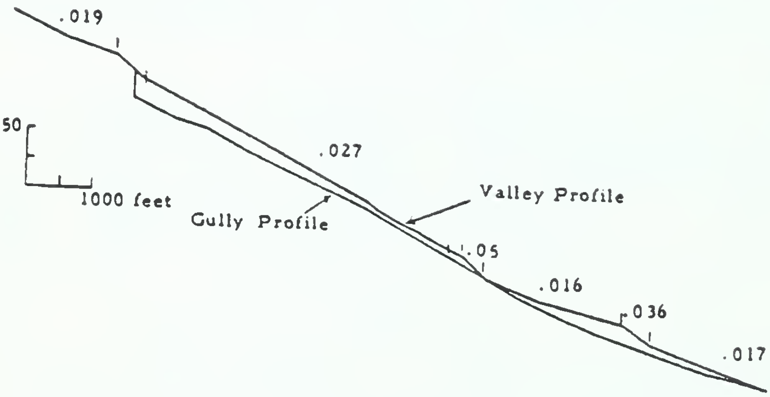

Figure 4-9. Longitudinal profile of Greasewood Creek, western Colorado. Note irregular valley-floor profile and discontinuous gullies cut below valley floor. Numbers above profile indicate slope of valley floor.

discontinuous gullies were observed, whereas 33 were stable or ungullied. All the valleys are within the semi-arid Piceance Creek basin in northwestern Colorado, and the land use, geologic, climatic and geomorphic conditions are very similar throughout the study area (Patton, 1973). Within each valley Patton (1973) identified those reaches which were gullied, or in the absence of gullies, the reach with steepest longitudinal slope. He surveyed the longitudinal slopes of these reaches and measured the drainage areas above them from topographic maps at a scale of 1:50,000. These data, together with Nebraska data published by Brice (1966, Figs. 2 and 3) serve as a basis for discussion.

The approach taken is essentially deductive, starting with a search for a possible threshold parameter, related to gully incision, which has a physical meaning. A reasonable choice is the average shear stress exerted by flow on the valley floor (Graf, 1979). This average shear stress is a function of the hydraulic radius of the flow, where hydraulic radius (R) is the ratio of cross-sectional area to wetted perimeter, the energy slope (Se) and the weight per unit volume of the water (Y). All are related by the equation:

$$\tau_o = \gamma R S_e \qquad (4\text{-}1)$$

For wide (shallow) flows, R can be replaced by the flow depth (d), and if the energy slope Se is approximated by the average valley slope S, equation (1) becomes:

$$\tau_o = \gamma \, dS \qquad (4\text{-}2)$$

It is necessary to relate flow depth to the drainage area. Using empirical relationships between flow depth (d) and water discharge Q, and similar relationships between water discharge of an event with a recurrence period of n years, (Qn) and drainage area (A) (Leopold et al, 1964, pp. 215, 251 and Table 7-5):

$$d = c_1 Q^f \qquad (0.36<f<0.45) \qquad (4\text{-}3)$$

$$Q_n = c_2 A^r \qquad (0.65<r<<0.80) \qquad (4\text{-}4)$$

c_1 and c_2 are constants. Substituting equation (4-4) into equation (4-3):

$$d_n = c_1 c_2{}^f A^{rf} \quad \text{(where c is a constant)} \qquad (4\text{-}5)$$

From the known range of the exponents f and r, their product rf is expected to be within the range:

$$0.23 << rf < 0.36 \quad \text{or, say: } 0.2 < rf < 0.4$$

Substituting equation (4-5) into equation (4-2), a relationship between average shear stress, drainage area and valley slope is defined. Since this relation is based on simplistic assumptions, τ_o is referred to as the shear-stress indicator. $\tau_o(n)$ then is an indicator of the actual shear stress $\tau_o(n)$, expected for a storm with a recurrence period of n years, and it is represented by:

$$\tau_{o(n)} = (c)A^{rf}S \qquad (4\text{-}6)$$

According to this equation, a line of equal values of the shear-stress indicator plots as a straight line on log-log paper. If valley slope S is plotted on the ordinate and drainage area on the abscissa, then the slope of such a line is equal to (rf).

Patton's data were replotted on a log-log paper with slope on the ordinate (Fig. 4-10). Following the reasoning of Patton and Schumm (1975), a straight line was drawn through the "lower most" points of the gullied valleys. This line is assumed to represent the threshold value τ_{th} of the shear-stress indicator, below which valley floors are stable. On Figure 4-10 the slope of the line is -0.26, so that rf = 0.26, which is within the limits deduced above for the rf exponent.

After the rf exponent was established from the data, equation (4-6) was used to calculate the values of $A^{0.26}S$, which is equal to $(1/c)\tau_o$, for all the data points. The lowermost value of $(1/c)\tau_o$ for those valleys which are gullied, was considered to represent the threshold value of $(1/c)\tau_{th}$.

Then, for each data-point, the value of τ_o/τ_{th} was calculated thereby eliminating c. This parameter is the relative shear-stress

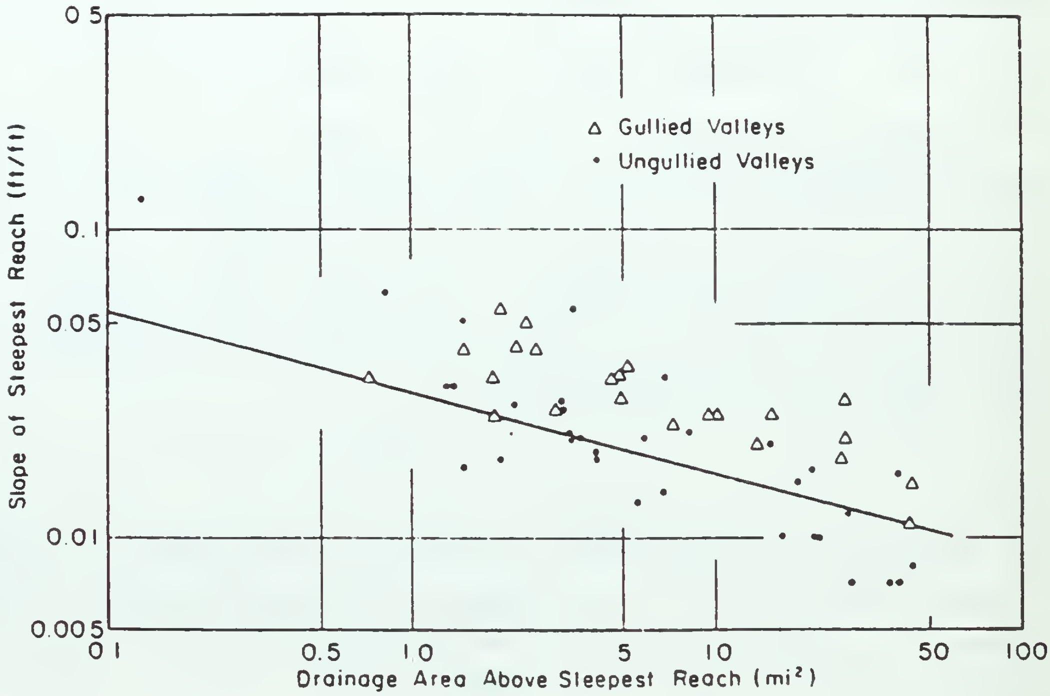

Figure 4-10. Replot of data from Figure 2-1 (Begin and Schumm, 1979).

indicator. Lines of equal τ_o/τ_{th} values are plotted on Figure 4-11 parallel to the basic line τ_o/τ_{th} = 1.0 which is the original line defined as τ_{th}.

Figure 4-11 shows that, although there are no gullied valleys in the area below τ_o/τ_{th} = 1.0 by definition, there are several ungullied valleys with τ_o/τ_{th} > 1.0. It is, of course, these reaches of valley floors that are potentially the most likely to be eroded in the future and from a land management point of view they are of high priority.

Examination of Figure 4-11 reveals that the ungullied valleys with small drainage areas tend to withstand higher values of the relative shear-stress indicator. This becomes clear when data for only ungullied valleys were plotted with Brice's (1966) data (Fig. 4-12). The τ_o/τ_{th} values are insensitive to change in drainage area for areas greater than about seven square miles. However, valleys with smaller drainage areas "survived" markedly increased values of τ_o/τ_{th}. A related observation was made by Patton and Schumm (1975) who pointed out that their discriminant function is not applicable to basins with small drainage areas. They explained this by suggesting that in small basins the aspect of the valley becomes a dominant factor. Other reasons may be differences in vegetation cover or shallowness of the alluvial mantle in low-order basins.

Although the sample used in this study is small, the results are encouraging. The method enables a planner to develop rational priorities of soil conservation measures based on an estimated probability of valley incision. However, the above numerical results pertain only to drainage basins with uniform geomorphic and hydrologic characteristics. Different values of the threshold shear-stress indicator are expected where geology, soils, climate and vegetation are different.

As an example, the data of Brice (1966, Fig. 2-3) were treated by the suggested method, and the results are shown on Figure 4-13. The sample is small but the similarity of the slope of the shear-stress indicator lines (negative exponent rf in equation (4-6)) in Figures 4-13 and 4-11 is noteworthy. Also, in both cases, valleys with small

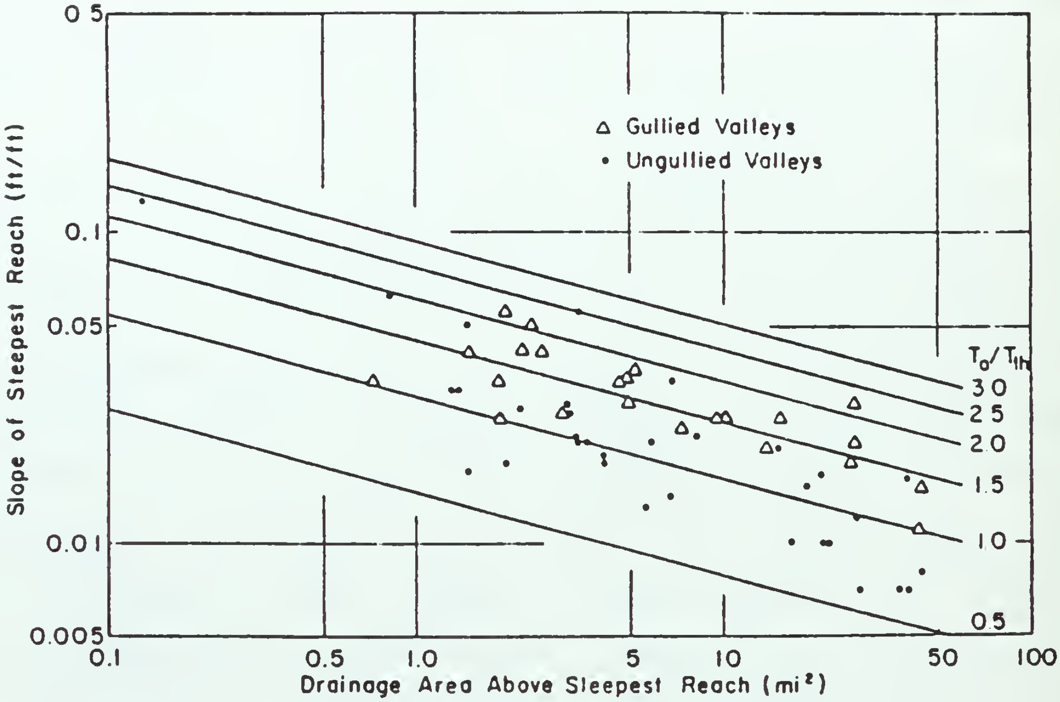

Figure 4-11. Plot of data from Figure 2-1 showing values of Relative Shear Stress Indicator. (Begin and Schumm, 1979)

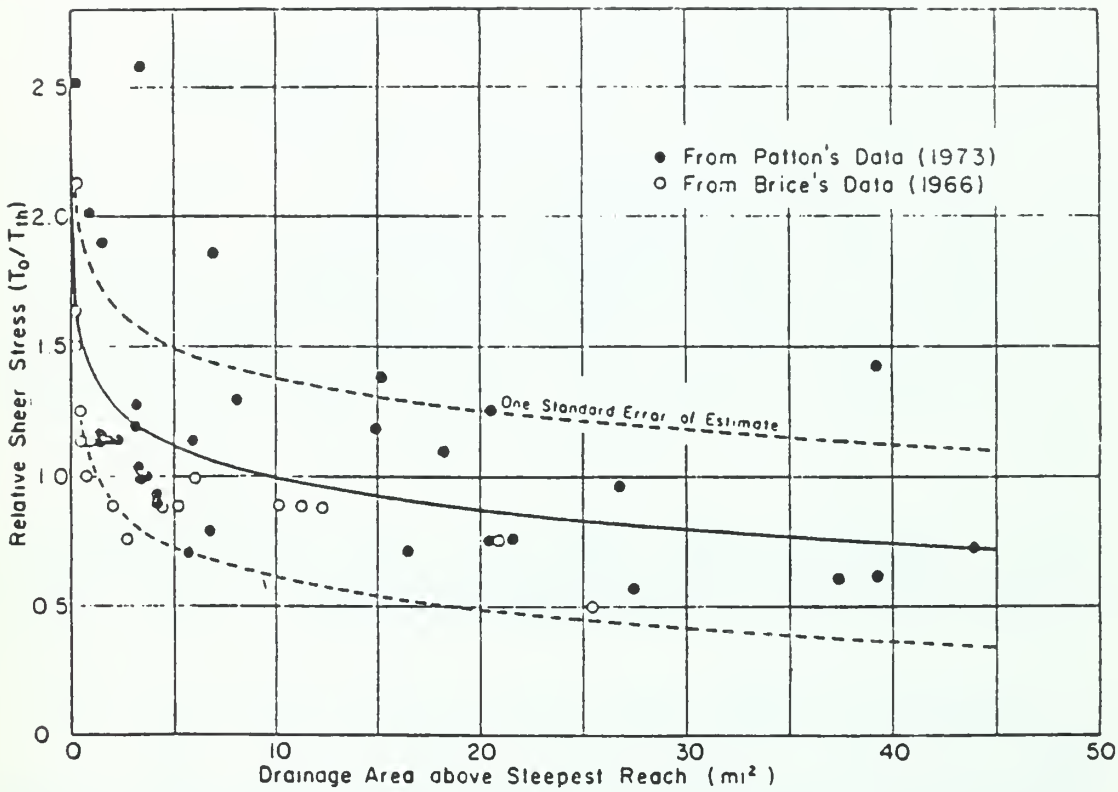

Figure 4-12. Relation between drainage area and Relative Shear Stress Indicator for ungullied valleys (Begin and Schumm, 1979).

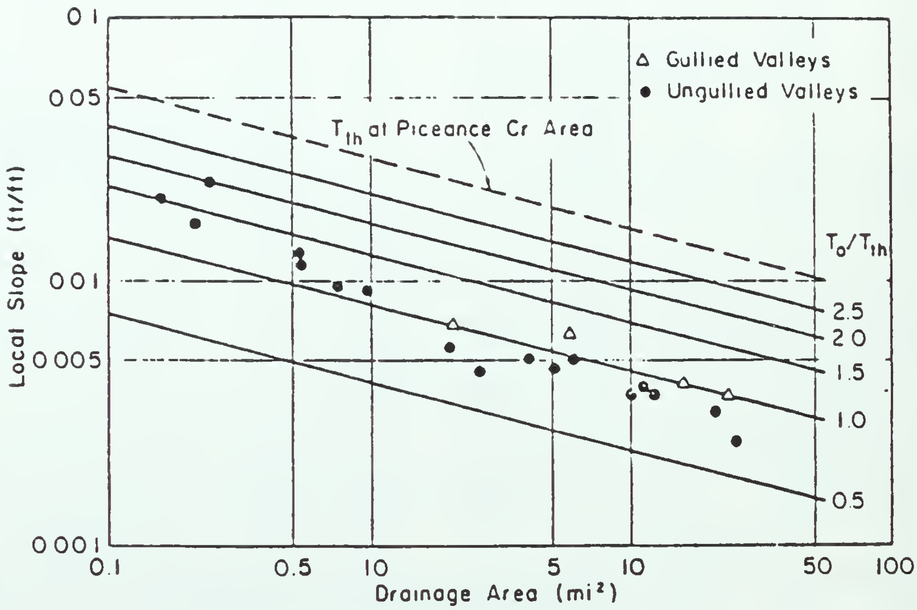

Figure 4-13. Drainage-area valley-slope plot showing Relative Shear Stress Indicator for Brice's Nebraska channels. (Begin and Schumm, 1979)

drainage areas tend to resist gullying while showing high values of the relative shear-stress indicator (see also Figure 4-12).

This study is based on the concept of geomorphic thresholds (Schumm, 1973, 1977). However, the results suggest a modified approach towards this concept. Figure 4-11 indicates that a threshold shear-stress indicator may be defined, below which gully incision does not occur. However, increased values of the relative shear-stress indicator do not imply that gullying will take place. Rather, it implies only an increase in the probability that a valley floor will be gullied. This seems reasonable in view of the basic stochastic mechanisms involved such as the temporal and spatial distribution of rainfall, the nonhomogeneity of soil and vegetation distribution, and measurement errors. Nevertheless, the equation of the threshold line of Figure 4.11 which is:

$$S = .008\ A^{-.26} \tag{4-7}$$

defines the critical slope for drainage areas up to about 50 square miles in areas of generally similar geology, climate, vegetation and land use.

Some of the gullied valleys had slopes greatly in excess of the threshold line. This indicates that the threshold line is not fixed in position, but it can shift vertically depending on climatic fluctuations and land-use changes. For example, any activity that increases flood peaks or average discharge will cause erosion of presently "stable" valley floors. The result will be a downward shift of the threshold line.

It is also important to stress that the definition of such a line (Equation 4-7) is only applicable to an area that is similar in all respects. For example, the Piceance Creek and Brice's threshold-line intercepts are very different (Fig. 4-13) reflecting different materials and climate.

Chalk Bluffs Area

In order to test the geomorphic approach to the identification of valley-floor stability the Chalk Bluffs and Pine Bluffs area of semi-arid northeastern Colorado were selected for further study of the threshold concept. As in the Piceance Creek area, the climate, geology, hydrology, basin morphometry, land use vegetation and soils vary little between basins. These variables, which would normally have to be incorporated in the analysis, were considered constants.

The Chalk Bluffs and Pine Bluffs form the pronounced escarpment at the boundary of the High Plains Physiographic Province and its Colorado Piedmont Section (Fenneman, 1931).

The discontinuous gullies in the Chalk Bluffs area are initiated on the pediment at the base of the bluffs. This is essentially a bedrock surface into which the channels have and are eroding.

The situation differs from the Piceance area in that the gullies are not confined within well-defined valleys. Once the drainage leaves the bluffs, the channels selected for study are within shallow bedrock depressions (valleys) on the piedmont. The bedrock occasionally locally confines the flow within these shallow valleys.

Continuous gullies exist further away from the bluffs, but they are formed by headward erosion in response to lowered base level and erosional activity of the South Platte River. The gullies develop in the siltstone and claystone of the Brule formation and its alluvium.

A total of 76 gullied or ungullied locations were studied in this area, 60 gullied sites and 16 ungullied sites (Bradley, 1980).

The examination of drainage basin and valley-floor characteristics indicates that in this area there are three major geomorphic factors that influence gully initiation. These are valley-floor longitudinal slope, valley-floor width, and drainage basin area.

Stable or ungullied valley reaches exhibit a characteristic concave-up longitudinal profile, but when local deposition or sediment storage occurs, the longitudinal profile of that reach becomes convex-up (Fig. 4-14). This is characteristic (Fig. 4-7, 4-9). Deposition continues until the locally over-steepened reach exceeds a critical threshold slope, when a runoff event of sufficient magnitude triggers gully initiation.

Gully initiation is also controlled by local changes of valley-floor width. A decrease of valley-floor width causes runoff to become more

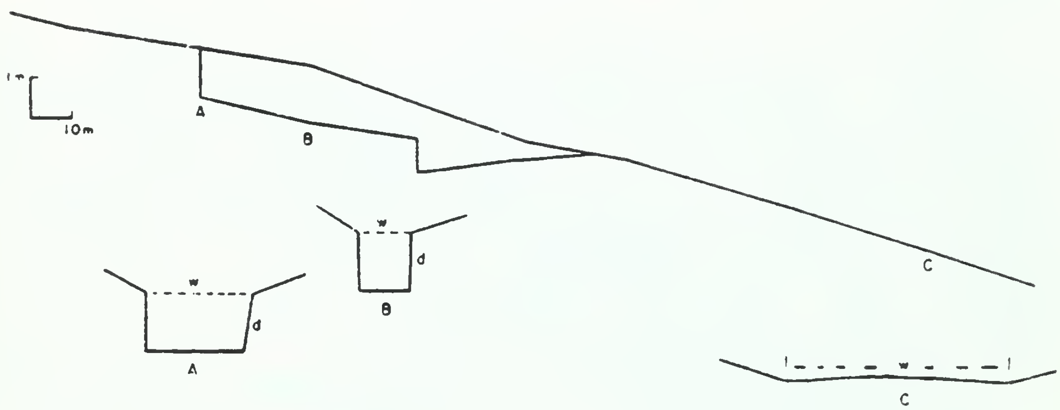

Figure 4-14. Longitudinal profile of Chalk Bluffs Valley and valley cross sections showing effect of valley-floor slope and width on gully location. (Bradley, 1980)

concentrated and therefore, it increases the depth and erosivity of the flow. It is important to note that in this area, valley floors are relatively flat, and flow events occupy the entire valley-floor. Therefore, valley-floor width can be used for flow width.

A combination of steep valley-floor slope and narrow valley-floor width creates a favorable condition for gully initiation. On Figure 4-14 valley-floor transverse and longitudinal profiles illustrate the combined effects of decreased flow width and increased slope upon gully initiation.

Although in the Chalk Bluffs area variations in climate, geology, vegetation, sediment and land use are relatively minor, nevertheless, a threshold slope could not be clearly identified. However, as noted above, the valleys are different, and at some locations the width of the valley is constricted by bedrock. In addition, the majority of the sampled valleys are within drainage basins smaller than the 7 square mile size that produced equivocal results in the Piceance basin study.

The procedure developed for the Piceance Creek area was modified to permit evaluation of the effects of valley width, and a ratio of valley slope to valley width is plotted against drainage area on Figure 4-15.

The relation of Figure 4-15 is less clear than that of Figures 2-1 or 4-10, but nevertheless, it is possible to draw two threshold lines on Figure 4-15. Above the upper line there are no ungullied locations, and below the lower line there are no gullied locations. A threshold zone has been identified based on the slope-width (S/W) ratio and drainage area. Only three ungullied locations plot below the zone, but many additional ungullied locations could have been found that would plot below the line. Nevertheless, the significant overlap suggests that the threshold line or zone has been shifted downward, perhaps as a result of the drought years of the 1930's or because of increased grazing pressure.

Figure 4-15 indicates that many of the presently stable or ungullied locations are susceptible to erosion (Bradley, 1980).

The shear-stress index of valley stability, as used earlier, must be modified because of the addition of valley width or flow width. In the

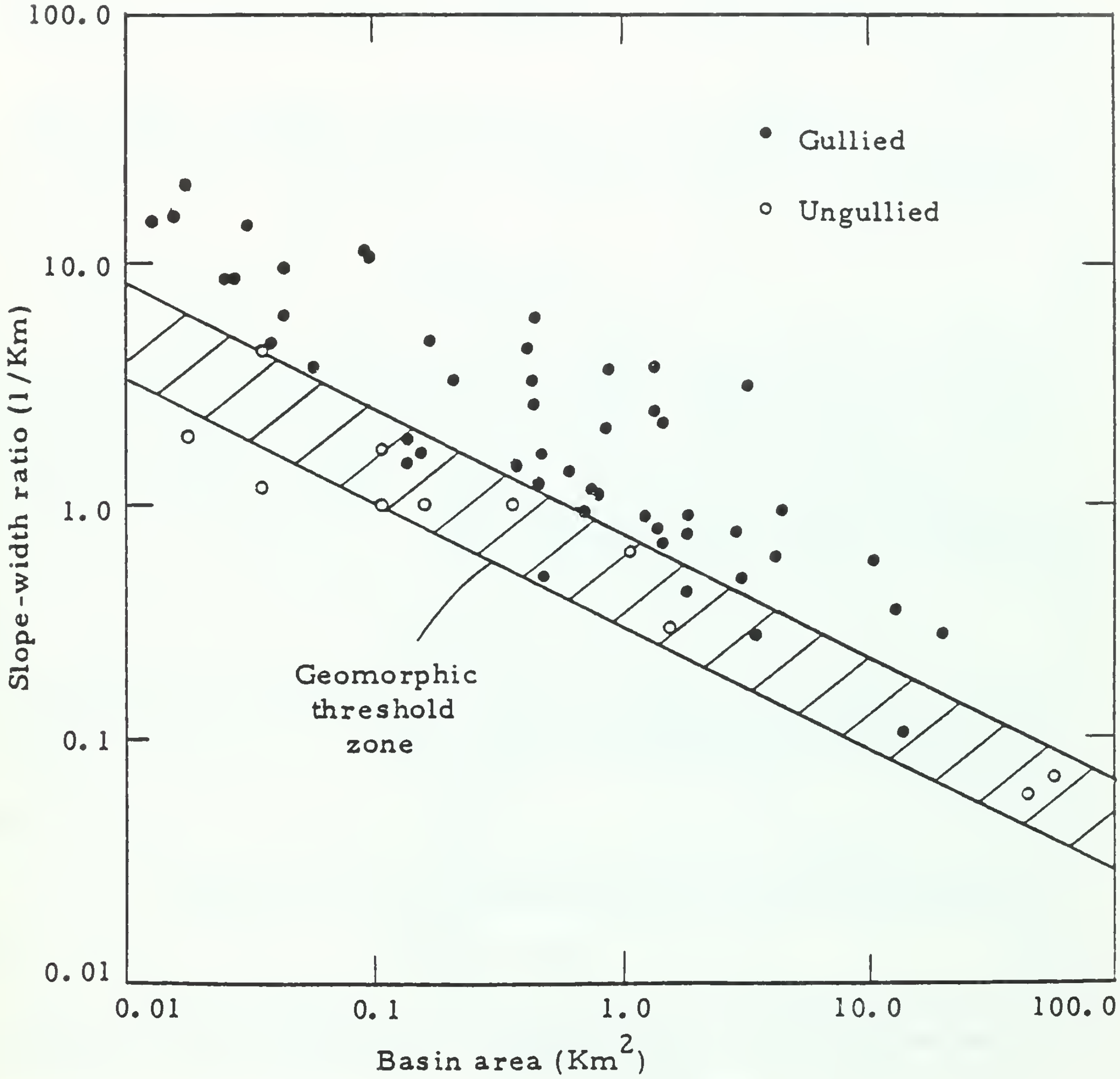

Figure 4-15. Plot of critical valley-floor slope to width ratio versus drainage basin area, illustrating the geomorphic threshold zone separating ungullied from gullied valley floors. (Bradley, 1980)

Chalk Bluffs area a stream power index was developed (Fig. 4-16). Bagnold (1966) defined stream power as "the rate of work performed by the transporting fluid or the rate of energy loss per unit length of stream."

The development of a relationship between stream power and geomorphic and hydraulic variables begins with the equation for shear stress (Equation 4-1):

$$\tau_0 = \gamma RS$$

where τ_0 is the shear stress at the water-channel interface, γ is the weight per unit volume of the water and it is considered a constant, R is the hydraulic radius of the flow, and S is the valley slope which is an approximation of the energy slope. Stream power is simply shear stress (τ_0) multiplied by velocity (V):

$$\tau_0 V = \gamma RSV \tag{4-9}$$

as indicated earlier hydraulic radius is the ratio of cross-section area (A_c) to wetted perimeter of a channel (P):

$$R = \frac{A_c}{P} \tag{4-10}$$

Substituting equation (4-10) into equation (4-9) yields:

$$\tau_0 V = \gamma \frac{A_c}{P} SV \tag{4-11}$$

Cross-sectional area (A_c) is related to discharge (Q) and velocity (V) by the continuity equation:

$$Q = A_c V \tag{4-12}$$

or

$$A_c = Q/V$$

Substituting equation (4-12) into equation (4-11) yields:

$$\tau_0 V = \gamma \frac{Q}{V} \frac{S}{P} V \tag{4-13}$$

or

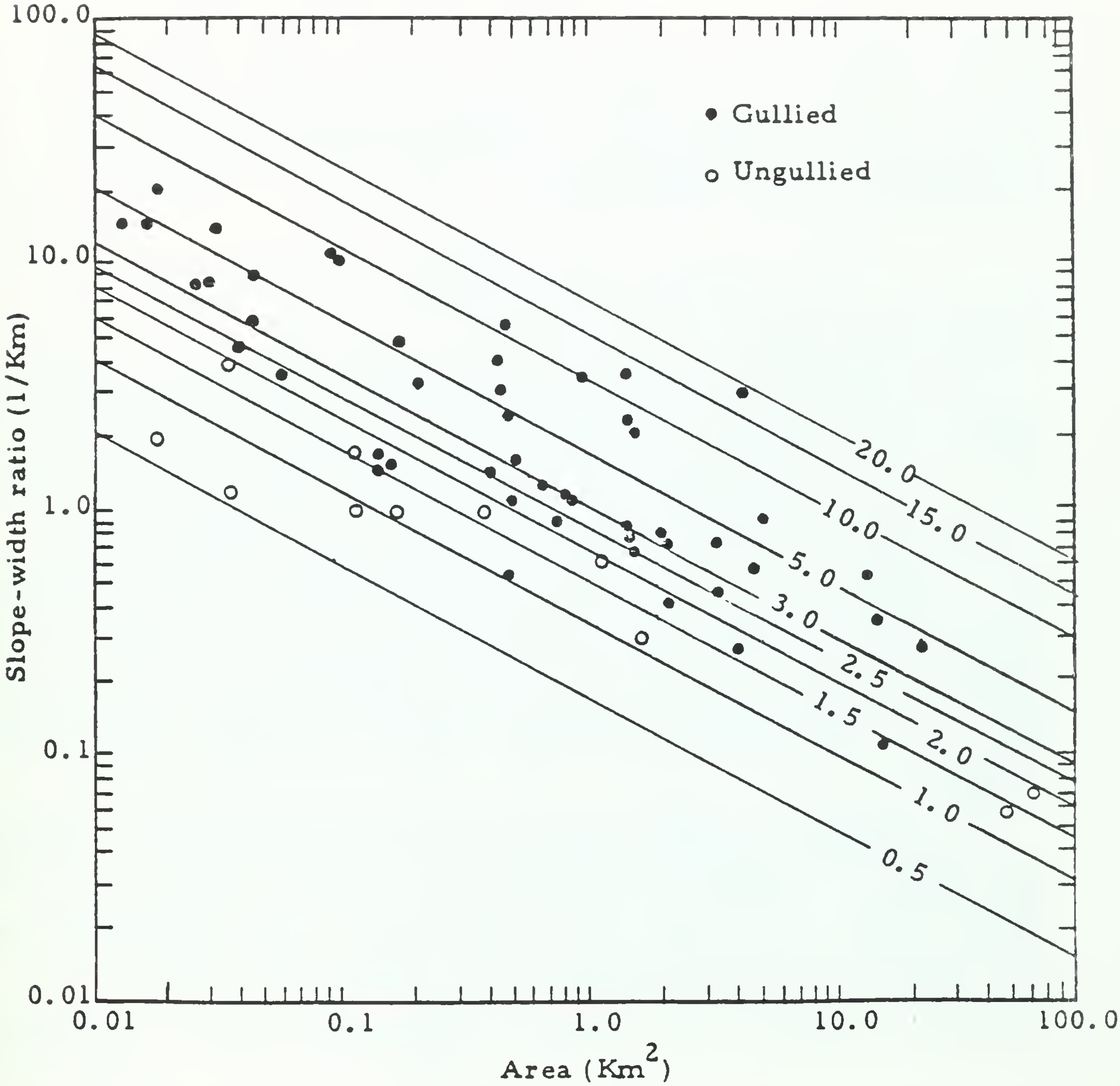

Figure 4-16. Geomorphic threshold or Stream Power Model. Lines show constant values of Relative Stream Power. Note the Geomorphic threshold zone of Figure 4-15 is between Relative Stream Powers of 1.0 and 2.0. (Bradley, 1980)

$$\tau_o V = \gamma Q \frac{S}{P} \tag{4-14}$$

Wetted perimeter (P) is best estimated by the valley-floor width (W). This is based on the assumption that the flow event occupies the entire valley floor. Discharge (Q) is best estimated by drainage basin area (A_b)(Burkham, 1966). Therefore:

$$\tau_o V = KA_b \frac{S}{W} \tag{4-15}$$

where K is a constant.

The advantage of defining the geomorphic threshold in terms of stream power instead of a simple critical-slope/critical-width ratio is that it allows consideration of all three variables of gully initiation. namely valley-floor slope, valley-floor width and drainage area.

There are two means by which the stream power geomorphic threshold can be exceeded and gully erosion initiated. This is achieved by either an increase of discharge or by an increase of valley-floor slope. Normally, channel width does not change, but artificial constriction of flow will effectively decrease the width and it will probably cause incision.

The distance a sample plots from the threshold line is related to the degree of stability or instability in terms of the critical stream power ($\tau_o V_{th}$). The stream power value ($\tau_o V_o$) of each location can then be compared to the critical stream power ($\tau_o V_{th}$) as a ratio:

$$\tau_o V_{rel} = \frac{\tau_o V_o}{\tau_o V_{th}} \tag{4-16}$$

to yield a value of relative stream power ($\tau_o V_{rel}$). Thus, a quantitative means for comparing each critical location's degree of stability is established. Lines of equal relative stream power ($\tau_o V_{rel}$) were calculated and plotted as lines parallel to the threshold stream power ($\tau_o V_{th}$) line in Fig. 4-16.

The equation of the lower line of the threshold zone is:

$$S/W = 0.35\ A^{-.55} \tag{4-17}$$

and hence, for any locations the critical slope-width ratio can be calculated (Bradley, 1980).

CONTINUOUS VALLEY-FLOOR GULLIES

Although the previous discussion has dealt with single gullies, as noted in Chapter 1, an entire drainage network may be involved in rejuvenation.

Brice's (1966) study of the gullies in Medicine Creek is unusual because it is a study of an entire drainage basin and the character of distribution of the gullies that lie within it. In most cases, gullies are studied individually; they are described in detail, but they are not considered in the context of the surrounding drainage basin. For this reason the Brice study is considered here in some detail.

Medicine Creek which is a tributary to the Republican River has a drainage area of 690 square miles (Fig. 4-17). The altitude of the basin ranges from 3,150 ft. to 2,400 ft. with a maximum local relief of about 200 ft. in the northern part of the basin. The drainage basin is underlain by in excess of 64 ft. of loess in the upper part of the basin and between 16 and 64 ft. in the lower one-third of the drainage basin, and the gullies cut into the loess. Annual precipitation varies greatly. The sixty-six year average between 1895 and 1960 was 21.5 in.

Medicine Creek basin is transitional between the tall grass and the short grass prairies of western United States. As noted above the loess is very thick and although there are occasional outcrops of other rocks they seem to have an insignificant effect on the erosional development of the area.

To provide some information on the distribution of gullies and headcuts within the Medicine Creek basin, Brice also shows the location of headcuts of the major valley-bottom gullies (Fig. 4-17). Most of the headcuts range in height from 10 to 25 feet in valleys of fourth or higher order. Advance of the headcuts during the period 1937 to 1952 ranged from a maximum of about 3400 ft. for gully B1 to a minimum of about 200 ft. for small channels (Table 4-1).

Note that the headcuts are most numerous in the central and lower part of the basin and headcuts in large valleys are commonly upstream

Figure 4-17. Nickpoints of major valley bottom gullies. Inset is map of location of Medicine Creek basin (Brice, 1966)

Table 4-1. Measurements relating to the size and topographic setting of major valley-bottom gullies. (Gully locations are indicated on Figure 4-17)

Gully	Drainage Area in 1952 (sq.mi.)	Width of Valley Bottom (ft)	Slope of Valley Bottom (ft per ft)	Height of Head Scarp in 1956 (ft)	Advance of Head Scarp 1937-52 (ft)	Estimated Year of Activation
B1	6.1	215	0.0056	24	900	1850
B2	5.6	210	.0055	22	750	?
B3	.8	120	.0105	20	610	?
B4	3.4	140	.0082	16	3,400	1900
B5	9.7	200	.0058	18	2,400	1920
B6	9.8	160	.0062	18	1,750	1920
B7	15.9	180	.00415	18	2,800	1910
B8	9.5	150	.0050	20	2,100	?
B9	1.4	120	.0125	23	2,100	1920

from the confluence of a major tributary. According to Brice, the greater number of scarps in the central and lower part of the basin can be explained by the relative narrowness of the valley bottoms, an analogy with the Chalk Bluffs area. Brice notes that the association of headcut and tributary junction is due to the fact that the main valley profile is generally steeper upstream from the confluence of the large tributary (Fig. 4-18). The steepening is probably due to deposition of a fan at the mouth of the tributary. Gully cross sections change downstream from the headcut. Typically gully depth decreases very gradually in a downstream direction, but gully width reaches a maximum within a thousand feet of the headcut and then decreases downstream (Fig. 4-19). The downstream widening effectively ceases when the width-depth ratio is in the range of 3.5 to 5. As the width increases the depth and velocity of the flowing water apparently is insufficient to remove the material that accumulates from bank slumping which prevents further widening of the channel (Table 4-1).

Brice's study was very significant for several reasons. He, like many others, indicated that there is an erosional evolution of the gullies. He suggests that the location of some of the headcuts and gullies can be associated with characteristics of the valley floor, which in turn leads to a means of predicting and identifying those portions of the valley floor that are most likely to contain gullies.

Additional studies indicate the problem that can be encountered during an attempt to develop relations like those presented on Figures 2-1 and 4-15. A study of gully evolution and distribution west of Trinidad, Colorado, failed to produce similar results. In this area coal mining, recreation activities and grazing have obscured the threshold because even the gentlest valley floor is incised, a result of flow concentration in roads and trails (Chamberlain, 1979). Hence, threshold conditions have been greatly altered by man's activities which overwhelm the innate geomorphic controls. The threshold approach must include the effects of man's activities if it is to be successful.

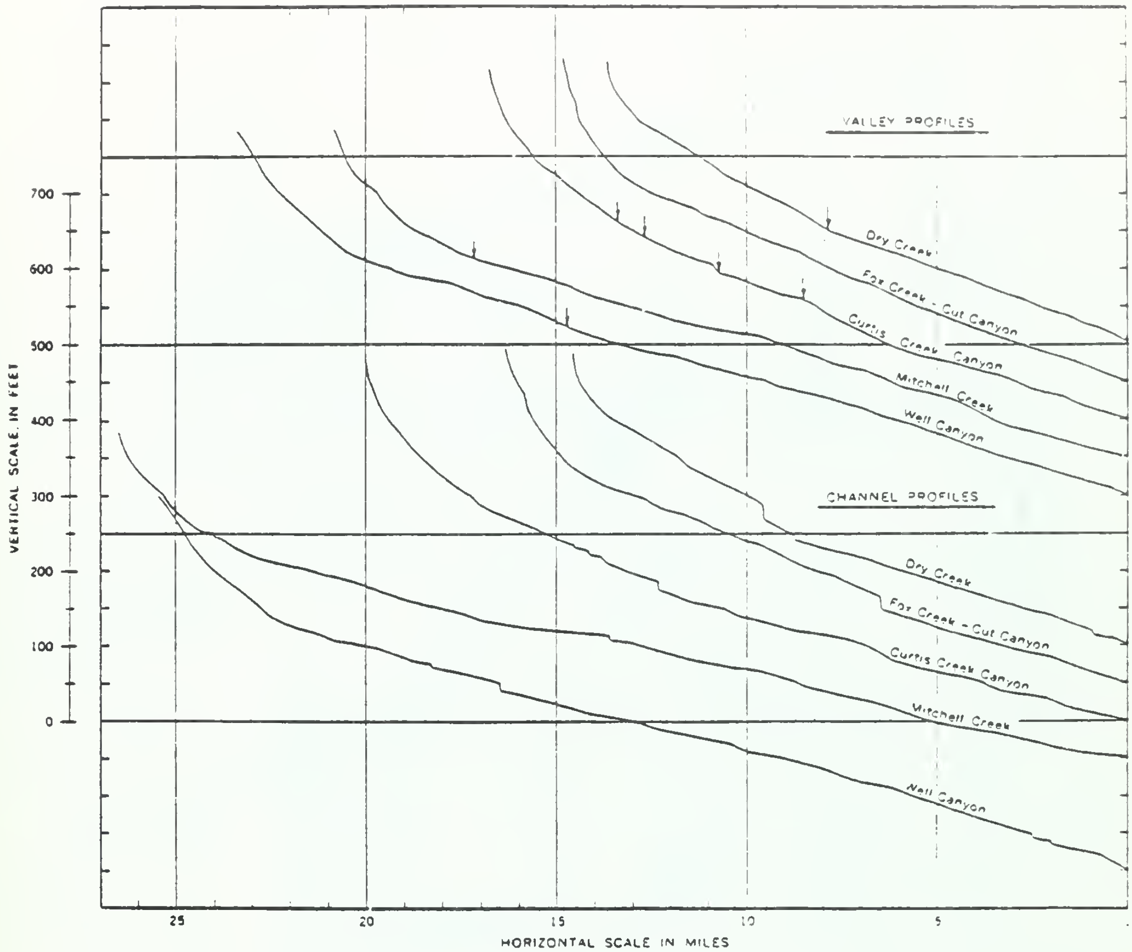

Figure 4-18. Long profiles of valley and channel for some major tributaries to Medicine Creek. A break in the channel profile represents a head cut, and an arrow on the valley profile indicates the point of confluence of a large tributary. (Brice, 1966)

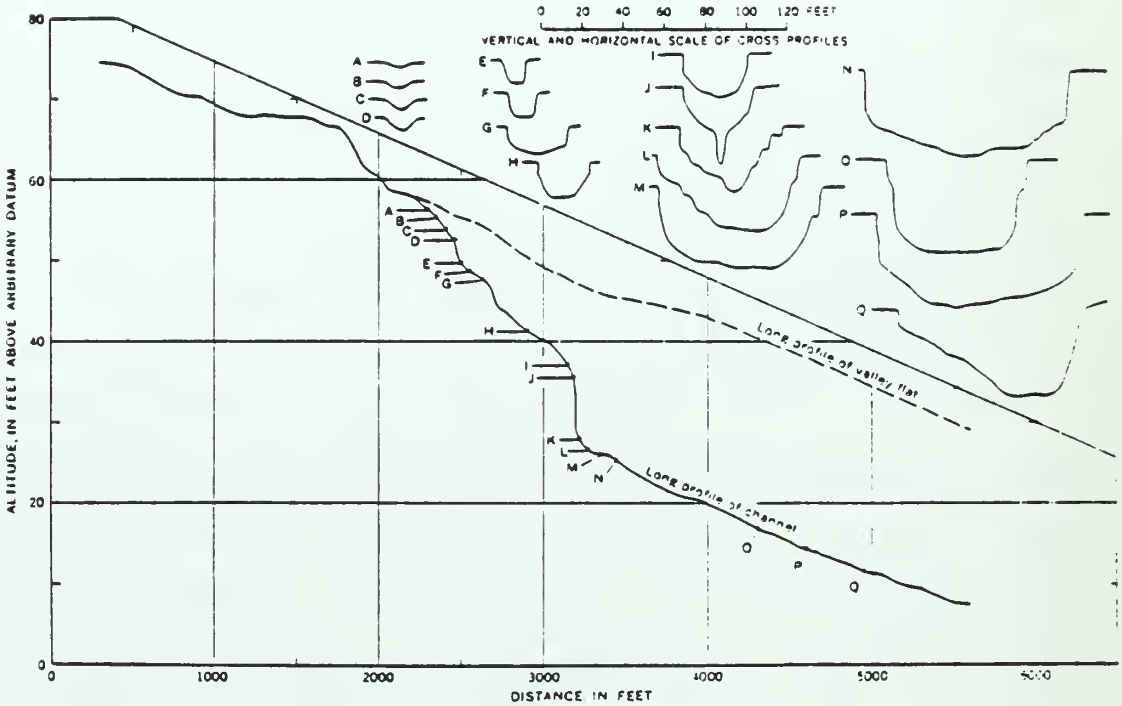

Figure 4-19. Long profile and cross profiles of a large valley-bottom gully in a tributary to Curtis Creek Canyon. Based on a field Survey 1956. (Brice, 1966)

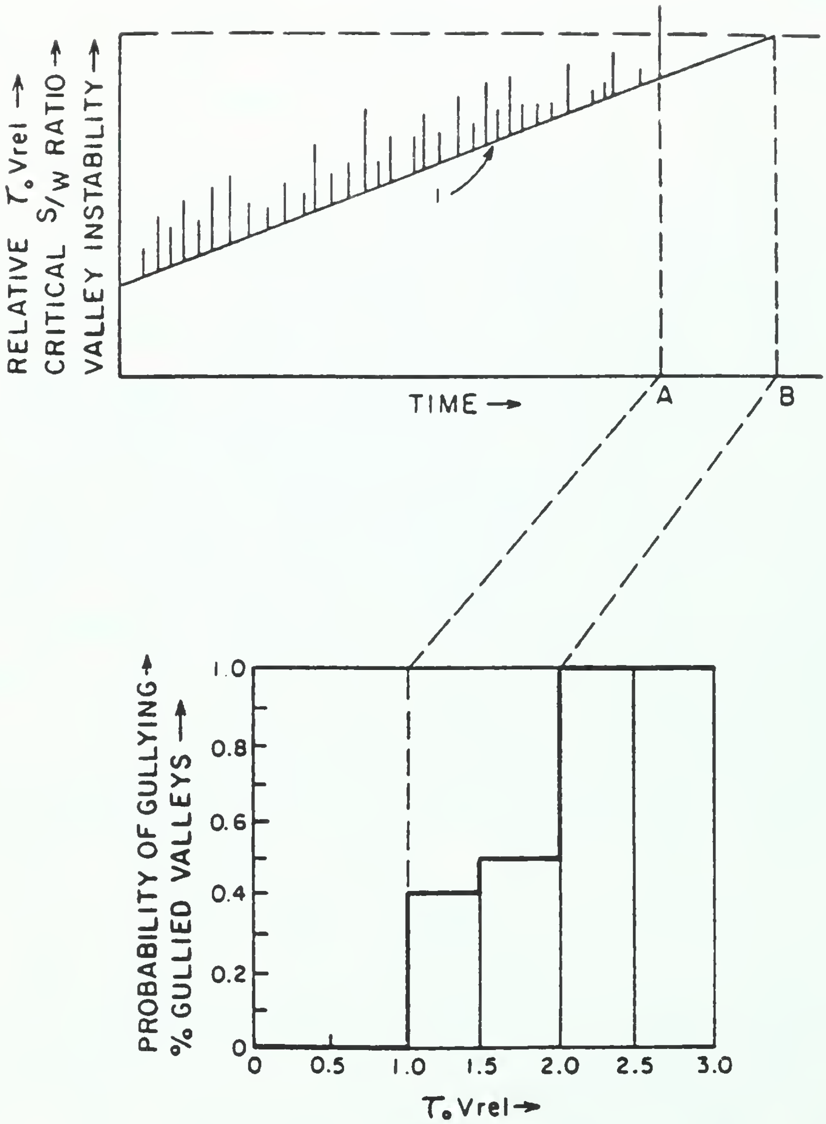

Figure 4-20. Change of valley-floor stability with time, as related to the probability of gullying for increased values of Relative Stream Power. (Bradley, 1980)

5. ENTRENCHED STREAMS

The term "entrenched stream" was introduced by Daniels (1966) to describe the conditions of the Willow River and other channels in western Iowa. These channels show the effect of straightening by channelization. This situation is typical of many areas of mid-western United States, where in an effort to reduce flood peaks and flooding of agricultural lands adjacent to the streams, the streams have been straightened with resulting degradation. A characteristic of the channelized streams is that erosion has been greatly accelerated by the increased gradient, decreased roughness and increased depth of flow.

For the most part, these channels are perennial. They carry at least small amounts of water during most of the year, and during seasons of high precipitation, large amounts of water are moved through the channels over a period of weeks and months. The channels cause serious problems both as a result of incision, which is a cause of rejuvenation of tributaries, increased upstream erosion and downstream sediment movement and deposition in reservoirs.

These channelized streams, the activity of which seem to be wholly the result of man's activities, appear to be analogous to the arroyos of the western United States. American Geological Institute Glossary (1972, p. 40) defines an arroyo as a "term applied in the arid and semi-arid regions of southwestern U.S. to the small, deep, flat-floored channel or gully of an ephemeral stream or of an intermittent stream usually with vertical or steeply cut banks of unconsolidated material at least 60 centimeters high; usually dry, but may be transformed into a temporary water course or short-lived torrent after heavy rains." Arroyos in the western United States are major incised channels extending for tens of km and they may even exceed a hundred km in length. Notable examples are the Rio Puerco and Chaco Wash in New Mexico and Polacca Wash, San Pedro Wash and San Simon Wash in Arizona. These are the best known examples of arroyos.

In the considerable literature relating to arroyos, Cooke and Reeves (1976) have recently reviewed the reasons for the development of arroyos, and their conclusion is that at least for many of the arroyos the initial cause of erosion was confinement of the flow by the development of roads and trails or other activities that have confined the flow and permitted the beginning of incision. However, similar arroyo development has occurred in the past due perhaps to climatic changes.

The more recent arroyo cutting undoubtedly is due to man's activities in the Southwest, but in addition, the channels of the valleys that have been trenched probably were geomorphically ready for this change in morphology. The conclusions of Schumm and Hadley (1957) relating to discontinuous gully formation on critical reaches of the valley also applies to the larger channels.

For the most part the arroyos are ephemeral stream channels that are subject to flash flooding. Rehabilitation of the arroyo is difficult, as the colonization of the channel by vegetation is hampered by the lack of water. In the eastern United States channelized streams for the most part are perennial, and vegetation can colonize the bed and banks much more readily. The discussion of entrenched streams, therefore, will deal with both the ephemeral channels of western United States and the channelized perennial streams of the midwest and eastern United States (Table 1-1).

LARGE ARROYOS

One of the first arroyos to be studied in detail was Polacca Wash in Arizona (Thornthwaite et al, 1942) in the Hopi and Navajo Indian Reservations. This study and that of Bryan (1928a and 1954) in the Rio Puerco Valley and in Chaco Canyon National Monument provides a basis for the examination of arroyo changes through time.

POLACCA WASH

Polacca Wash rises on the northeastern rim of Black Mesa and its course extends for essentially a hundred miles to its junction with the Little Colorado River near Luppe, Arizona (Thornthwaite et al, 1942). The upper half of the wash lies in canyons deeply carved in the mesa, but the lower half of the drainage basin is more open, and the wash traverses a broad plain. Other similar and more or less parallel washes include Moenkopi Wash, Dinnebito Wash, Orabi Wash, and Jeditto Wash. These are collectively known as the Tusayan Washes.

Precipitation in the Polacca drainage ranges from 7 inches per year at Luppe to about 12 in. at Keams Canyon. Precipitation is higher farther to the north on the mesa.

The geology of the drainage basin consists essentially of relatively flat-lying sandstones and shales, highly erodible materials that produce large quantities of sediment. Flow in the wash is intermittent and there is rarely a continuous stream of water from the headwaters to the mouth. The most striking feature of the present landscape of the Polacca Wash area is the great arroyo whose deep channel cuts the valley for a distance of 120 km. Twisting along the center of the valley with its bed 3 to 15 m below the valley floor, Polacca Wash widened its channel by meandering, and it has produced millions of tons of sediment. Polacca Wash was formed by the merging of at least five large discontinuous gullies which formed a continuous channel reaching from the headwaters to the Little Colorado River.

According to Thornthwaite et al (1942) there are stages of arroyo growth very similar to those described for the gullies of South Carolina as follows: (1) initiation; (2) enlargement by headward elongation; (3) healing, by reduction of slope of the gully walls and establishment of vegetation; and (4) stabilization, revegetation and possible eventual filling or obliteration. Unlike the characteristic decline of gully walls in more humid regions lateral cutting maintains steep valley walls, and the lateral cutting tends to enlarge the arroyo.

Figure 5-1 shows the effect of lateral migration and growth of meanders on the arroyo. In addition, it shows that there are five terraces in the arroyo. This is another example of the episodic erosion previously described for the Douglas Creek, Colorado area (Fig. 2-4). In these dry regions, where very large quantities of sediment are in transport, episodic erosion is expectable. The development of the main Polacca Wash gully presumably followed a sequence very similar to that of the small discontinuous gullies that have developed elsewhere. It

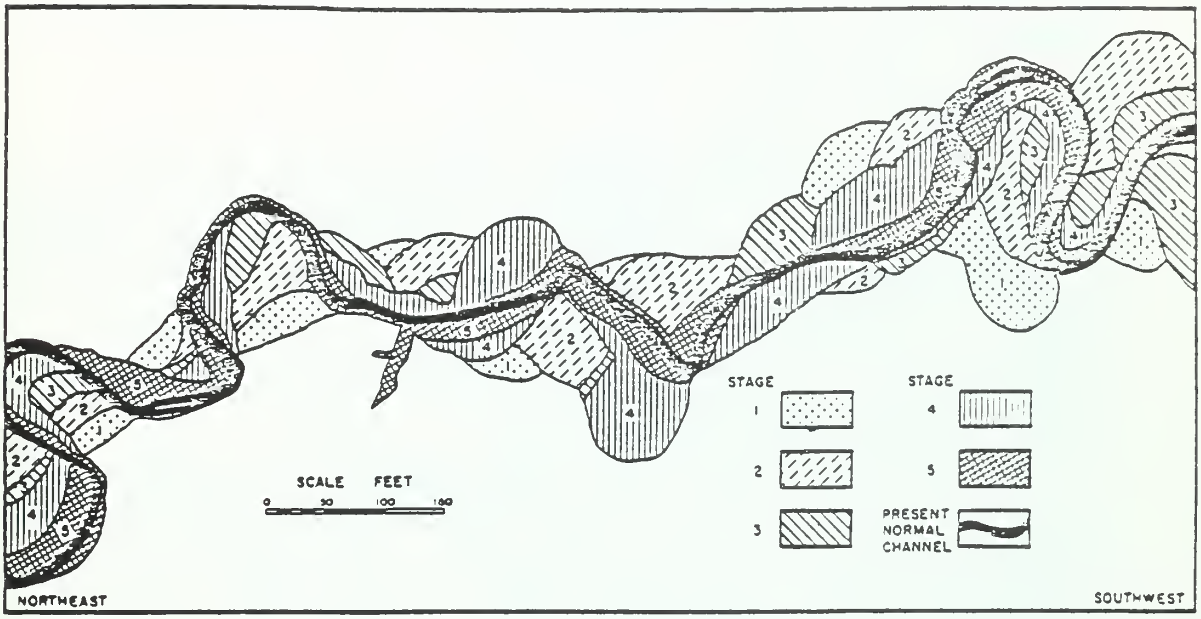

Figure 5-1. Segment of the Polacca Gully about a mile south of Polacca village. The numbered patterns indicate minor erosion levels occupied temporarily by the channel as it cut downward and migrated down valley.(Thornthwaite, et al., 1942)

seems inevitable that the downstream reaches of Polacca Wash will eventually aggrade. The runoff and large sediment supply from the tributaries in the uplands move large quantities of sediment into the channel system. As the floods move downstream, however, the water is lost both into the banks and bed of the channel, and the resulting deposition forms a channel plug, or fan.

RIO PUERCO

One of the major tributaries to the Rio Grande in New Mexico is the Rio Puerco. Numerous investigators have been interested in the changes of the Rio Puerco during the last century since its incision in the late 1880's. A considerable controversy exists as the cause of this incision, but for the purpose of this discussion it is only necessary to consider channel changes through time.

The Rio Puerco provides a classic example of entrenched channel development, which occurred throughout the Southwest at the end of the nineteenth century. It drains an area of 7,340 square miles in north central New Mexico, and it is the largest tributary to the Rio Grande in the middle Rio Grande valley (Fig. 5-2). From its headwaters in the Nacimiento Mountains near Cuba to its confluence with the Rio Grande near Bernardo, Rio Puerco is about 190 km long.

Rio Puerco flows southward across the structural boundary of the Colorado Plateau into the Rio Grande Depression (Wright, 1946). Its headwaters are underlain by crystalline rocks of the Nacimiento Mountains; however, the major portion of the valley is underlain by sandstone and shale of Tertiary and Cretaceous age.

The climate of the Rio Puerco area is semi-arid. Rainfall ranges from 14 inches per year at Cuba, to 6 inches per year at Bernardo near the confluence with the Rio Grande. The rainfall occurs in the months of August through October in the form of convective showers.

Between 1887 and 1928 an estimated 394, 882 acre-feet of sediment or 33 million tons per year of sediment was transported out of the Rio Puerco valley (Bryan and Post, 1927).

Before about 1885 the channel of Rio Puerco and the channels of its tributaries were small and inconspicuous. Bryan and Post (1927)

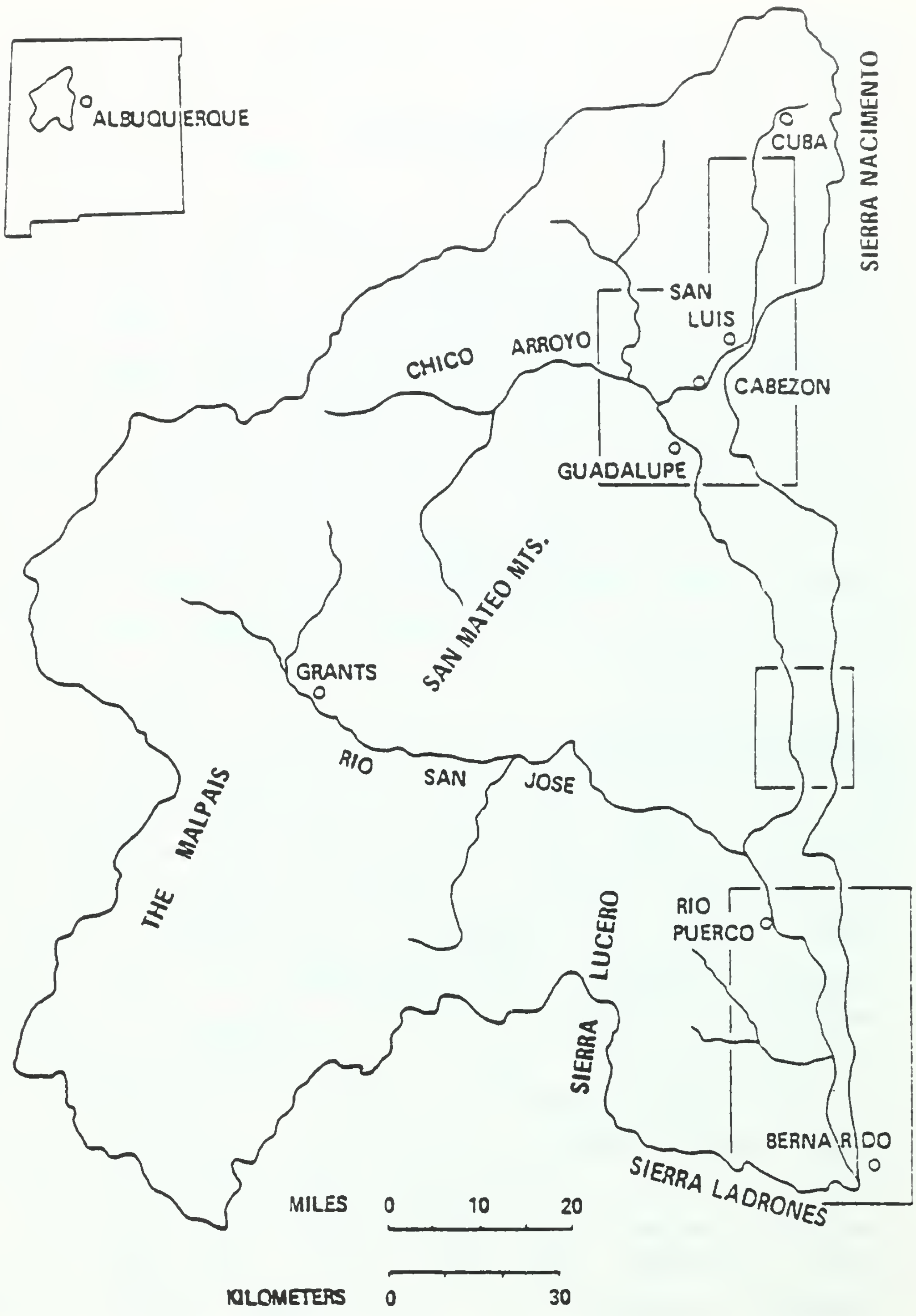

Figure 5-2. Rio Puerco, New Mexico; watershed and study areas (Elliott, 1979) showing location of cross section's discussed in text.

estimated the volume of the main channel and its tributaries to have been 22,500 acre feet ($2.78 \times 10^7 m^3$) before incision and 417,000 acre feet ($5.15 \times 10^8 m^3$) afterward, therefore, net erosion was 394,000 acre feet ($4.87 \times 10^8 m^3$) in 42 years, or an average of 9400 acre feet ($1.16 \times 10^7 m^3$) per year.

Accelerated erosion also has had an effect on water quality. Wagner (1963) notes a suspended sediment concentration of over 100,000 parts per million (ppm) in Rio Puerco, of which 50 percent was clay-sized particles. The capacity for transport of sand is tremendously increased with a high concentration of fine sediment. Nordin (1963) documented a 30 m^3 per second (1080 cfs) flow having a sediment concentration of 327,000 ppm with 40 percent of the sediment of sand size. In addition, he states that Rio Puerco, because of its high percentage of suspended sediment (40-50%) may transport as much as ten times more sand than the Rio Grande, given the same discharge and provided that such a quantity of sand is available.

Historical Background

Prior to 1885, the valley of Rio Puerco was stable with the exception of several discontinuous gullies (Bryan, 1928a). Portions of the valley were irrigated by diverting floodwaters across the valley floor. Because the inhabitants were dependent on this form of floodwater irrigation, Bryan (1928b, 1929) was able to document the deepening of the channel of Rio Puerco by the date of the abandonment of the communities in the valley. He stated that the town of Los Cerros, 34 miles above the mouth, was abandoned in 1887 and that the gullied channel reached Cabezon, 110 miles above the mouth, between 1885 and 1892 (Fig. 5-3). This evidence suggests that the formation of the continuous channel of the Rio Puerco was not the result of a single headcut, but it was the result of the nearly simultaneous deepening and subsequent coalescence of the previously existing discontinuous channels.

Bryan (1928a) suggested that poor land management practices preceding the period of destructive erosion was a "trigger pull" which

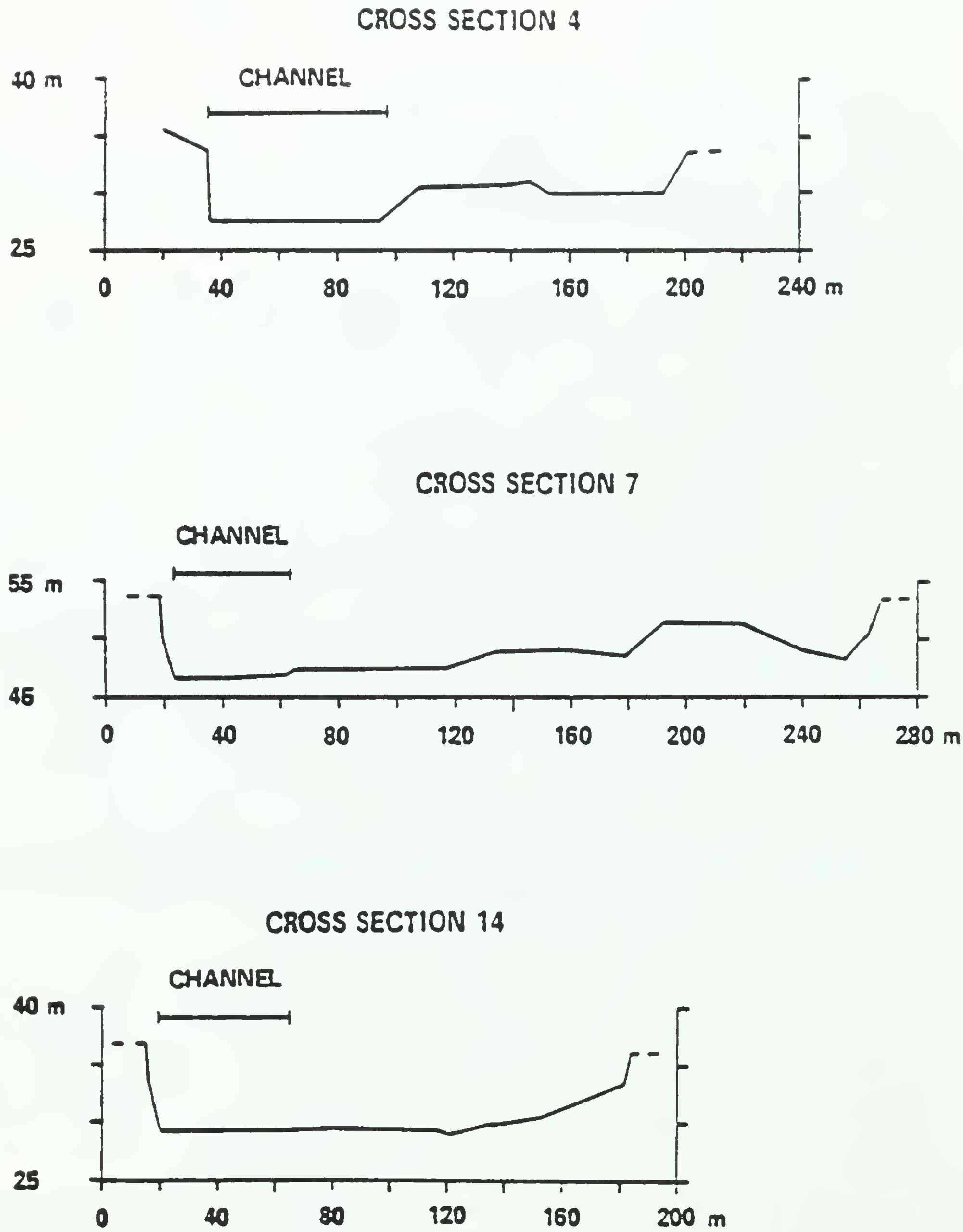

Figure 5-3. Rio Puerco cross sections 4, 7 and 14, Type 1 reaches. (Elliott, 1979). For locations see Fig. 5-2.

coincided with a natural climatic change to cause the gullying. The entire length of the river has now incised. The valley floor is no longer the floodplain, and rejuvenation of tributaries and widening of the main channel is in progress. At present, the arroyo exhibits different phases of enlargement. In the upper reaches and where major sediment carrying tributaries enter, the arroyo is wide with steep banks. The channel within the arroyo is wide and shallow, and its banks are easily erodible and poorly defined. Here the Rio Puerco carries much sediment.

Vegetation is generally scarce or young being eradicated periodically by floods. Braided channels and dissected transverse bars characterize these sediment choked areas during periods of low flow. Channel shifting is common during high discharge, and most of the arroyo is being actively eroded. Reaches of the arroyo that exhibit such a morphology and that are being rapidly enlarged by lateral erosion are referred to by Elliott (1979) as "Type-1" reaches (Fig. 5-3).

Downstream a broad arroyo having a width often in excess of 400 meters replaces the Type-1 arroyo, and it is referred to as a Type-2 arroyo (Elliott, 1979). The well-defined channel flowing through this inner-valley meanders with occasional shifts or avulsions and in most places, it is flanked by low terraces and dense stands of salt cedar. Vegetation on the adjacent flood plain and lower terraces is less dense, but it is in striking contrast to the barren appearance of the highest terrace, the pre-incision valley floor.

Notable differences in channel geometry and pattern exist between Type-1 and Type-2 river segments (Figs. 5-3, 5-4). Type-1 reaches are generally broader, shallower and steeper than Type-2 reaches, and they also have a larger bend radius of curvature and a lower percentage of silt and clay in the channel perimeter (Table 5-1).

As observed in the 1935 aerial photographs, most of the Rio Puerco was characterized by Type-1 conditions; it was broad, sandy and braided in appearance. When the next aerial photographs were taken in the mid 1950's much of the river exhibited the same conditions, but in some lower reaches the channel was much narrower than in the 1935 photos, and

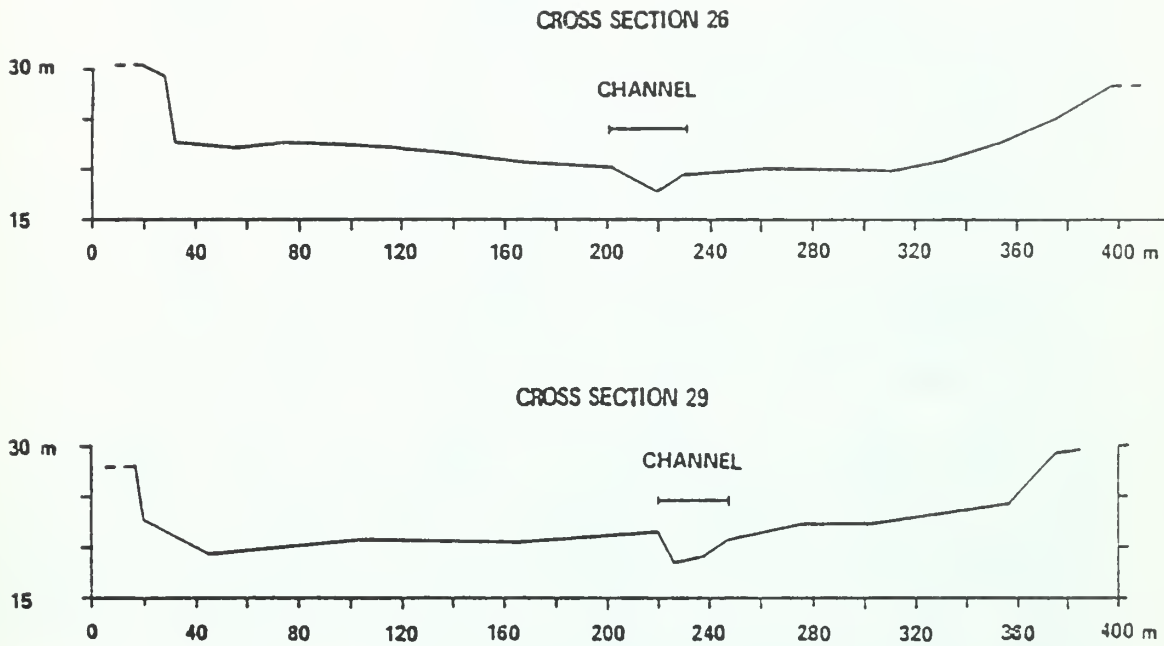

Figure 5-4. Rio Puerco cross sections 26 and 29, Type 2 reaches. (Elliott, 1979). For locations see Fig. 5-2.

Table 5-1 Rio Puerco Channel Characteristics.

Variable		Mean Type-1 Channel	Mean Type-2 Channel
Width-depth ratio	(W/D)	78.3	12.2
Gradient	S (m/m)	.0032	.0014
Silt-clay in channel	M (%)	23.7	44.9
Radius of curvature	r_m (m)	97.7	69.6
Sinuosity	P	1.0	1.5
Valley slope	S_v (m/m)	.0032	.0019

vegetation appears to have become established along banks and on developing point bars. In two decades, a Type-2 channel begun to supplant Type-1 conditions in the lower reaches, an indication that arroyo evolution was proceeding rapidly (Fig. 5-4).

Evolution of Rio Puerco Arroyo

The propensity for arroyos to broaden after the initial entrenchment has been addressed by several investigators. Following initiation, the maximum depth of a youthful arroyo is quickly achieved, but thereafter little additional deepening occurs despite the presence of easily erodible material (Leopold and Miller, 1956). Removal of valley fill continues at a rapid rate via lateral erosion of arroyo walls rather than by continued downcutting. Cooke and Reeves (1976) and Leighly (1936) have emphasized the general tendency for creation of a new channel and flood plain that is capable of accommodating flood events of widely differing magnitudes and frequencies.

It should be stressed that arroyo widening, owing to shifting of the channel within the arroyo, should not be confused with widening of the channel itself. When first entrenched, a channel may flow from scarp to scarp, but with continued channel shifting and lateral erosion of valley fill, the breadth of the arroyo will come to greatly exceed the width associated with flow in the channel.

Bryan and Post (1927) suggested that as the main channel and tributaries continue to erode and shift laterally, the arroyo of the Rio Puerco would one day evolve into a broad valley that would be similar in appearance to the pre-incision valley but adjusted to a lower base level. In a more recent study of the Rio Puerco, Elliott (1979) and Patton (1973) have found evidence that corroborates this hypothesis of valley formation. Along lower reaches of the river, lateral erosion of valley fill has widened the arroyo and point bar formation and overbank deposition has created a new flood plain within the broadened arroyo.

Long-term changes in many large arroyos follow the sequence developed by Elliott for Rio Puerco (Figs. 5-5, 5-6). Prior to rejuvenation, streams flow through and periodically inundate their flood plains (Figs. 5-5a, 5-6a). Incision follows, triggered by any combination of natural or cultural factors, and the entire discharge is confined to a narrow and deep trench (Figs. 5-5b, 5-6b). Highly erosive flows attack the walls of the trench and widening of the arroyo begins (Figs. 5-5c, 5-6c). Large quantities of sediment, derived from the rejuvenation of tributaries upstream and from the collapse of arroyo walls, overburden the main channel causing the stream to braid and shift laterally, further widening the arroyo (Figs. 5-5d, 5-6d).

As the arroyo is broadened, inner-floodplain construction begins with deposition of sediment within and adjacent to the channel. Eventually, sediment production decreases resulting in a change of channel character from braided to narrow and sinuous. Floodplain construction within the arroyo continues by overbank deposition, the building of point bars and the establishment of vegetation on these sediments (Figs. 5-5e, 5-6e). The evolution of the arroyo, now valley-like in proportions, proceeds as terrace scarps are reduced in angle by hillslope-erosion processes and as vegetation on the inner-flood plain becomes permanent and diversified.

An important distinction should be made between aggradation in small and large ephemeral drainages. Aggradation in discontinuous gullies and

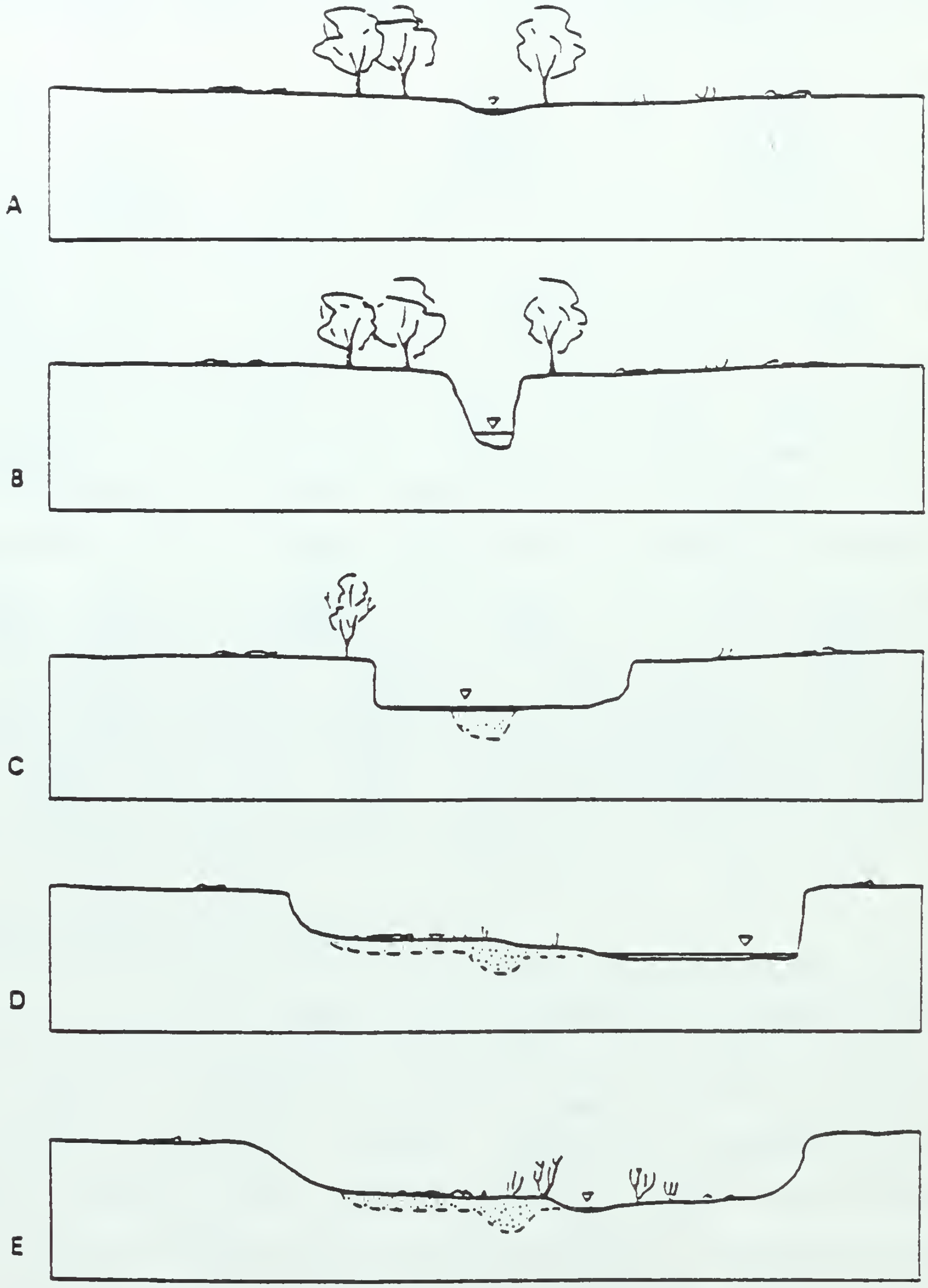

Figure 5-5. Hypothetical sequence of arroyo evolution. (Elliott, 1979)

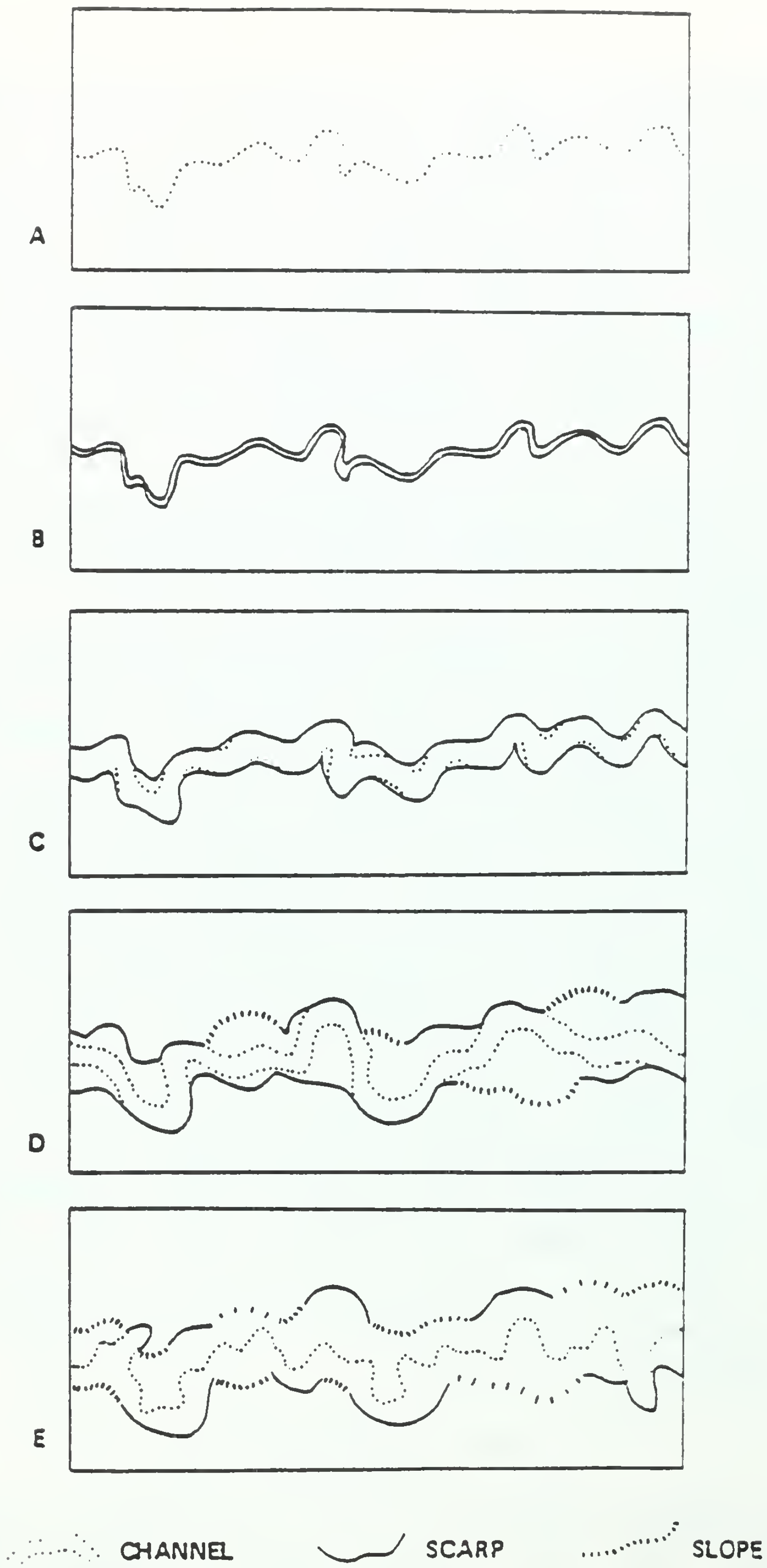

Figure 5-6. Hypothetical sequence of arroyo plan geometry evolution (Elliott, 1979)

small arroyos will often result in complete filling of the channel as the cycle of erosion progresses (Schumm and Hadley, 1957). But in large arroyos, such as that of the Rio Puerco, the magnitude of runoff events is great enough to offset any tendency for the channel to aggrade to its former level. In this respect, channels flowing through large arroyos behave more like alluvial rivers than ephemeral streams (Patton, 1973).

In summary, the behavior of large arroyos during a cycle of erosion is different than that of smaller-scale, discontinuous arroyos and gullies. While the gullying of small channels may soon be followed by aggradation to the level of the former valley floor, there exists a tendency for the large arroyos to become valley-like over a period of several decades.

In contrast to the laterally eroding upstream reaches, the stable appearance of the lower Rio Puerco implies that some optimal valley dimensions must exist. Although no empirical relations have been established for arroyo width, the inner-valley of the lower Rio Puerco where lateral erosion of the old valley alluvium has abated to a large degree, often exceeds 350 meters. Upper reaches of the main channel flow through an arroyo whose width is usually well below 200 meters, and it is along these segments that arroyo widening is most active.

This type of major incised channel, has formed a new valley, floodplain, and channel following incision. It is different from the smaller gullies and ephemeral streams that plug, backfill and obliterate themselves. The Rio Puerco may provide a model for incised channel evolution in humid regions. With deep incision and abundant runoff, widening of the channel and development of a new floodplain is inevitable unless the channel is fixed in position by structural means. However the widening may be inhibited by resistant sediments or by structural control.

EFFECT OF SEDIMENT TYPE

Three additional examples of arroyo change based on investigations carried out in 1957, 1971 and 1979 (Schumm, 1961; Patton and Schumm, 1981), considered three ephemeral-stream channels are located in South

Dakota (Sage Creek), Nebraska (Sand Creek) and New Mexico (Arroyo Calabasas) (Fig. 5-7). In addition to providing a record of channel changes during the 14-year and 22-year periods (Arroyo Calabasas was not visited in 1979), the three areas are characterized by different types of sediment and, therefore, the behavior and the characteristics and morphology of the channels are different. The study provides a means of evaluating the effect of sediment type on channel morphology and dynamics. This is a variable that cannot be evaluated by studies of single channels.

The three channels drain source areas that have high sediment yields ranging from predominantly sand (Arroyo Calabasas) to a mixture of sand, silt and clay (Sand Creek) to largely silt and clay (Sage Creek).

Channel morphology changed significantly between 1957 and 1979 (Patton and Schumm, 1981). Erosion occurred through nickpoint recession and bank collapse, but erosional reaches are separated by aggrading or stable-channel reaches. In general, sediment that is eroded, as the nickpoint recedes upstream, is trapped in the widened channel downstream. In this manner sediment is transported episodically out of these basins during a series of cut and fill cycles. The manner by which the channels aggrade and the morphology of the aggraded stable channels is controlled by the sediment type. The wide shallow channel of Arroyo Calabasas is filled by vertical accretion of sand-size sediment. The narrow and deep channels of Sage Creek and Sand Creek are created by the lateral accretion of cohesive fine-grained sediment.

In the Sage Creek area there is a high percentage of silt and clay in the stable channel cross sections because Sage Creek drains an area underlain primarily by siltstone and claystone. Deposition took place initially around slump blocks that had fallen into the channel when the width-depth ratio was about 5. The highly cohesive sediment is not immediately removed by flowing water, and when deposition continued it was greatest on the channel sides initially causing a narrowing of the channel at the bottom (Fig. 5-8). In Sand Creek, the silt clay in the stable channel sections was about 20%, which reflected the sandy nature of the channels exposed in the drainage area.

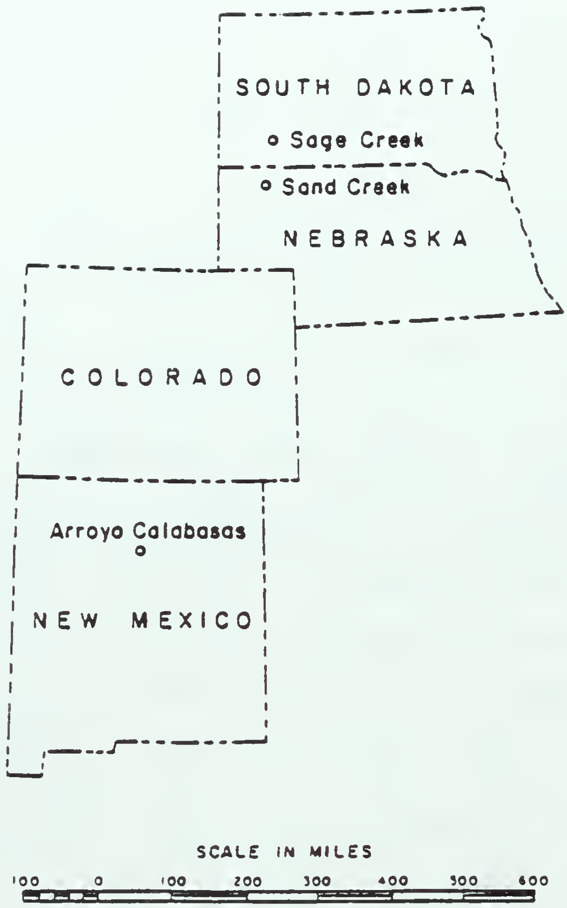

Figure 5-7. Location map of Sage and Sand Creek and Arroyo Calabasas.

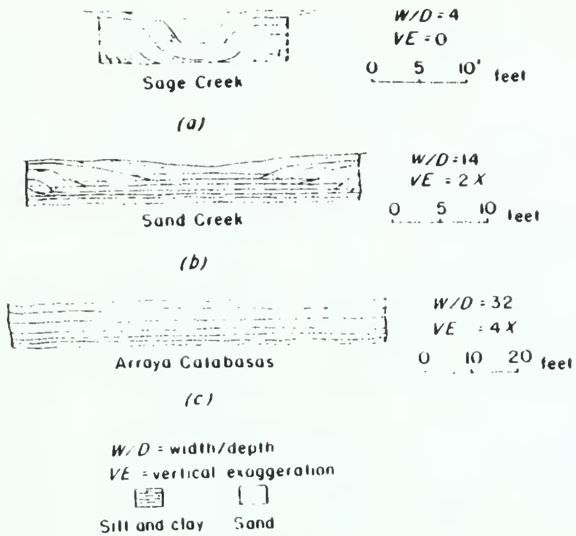

Figure 5-8. Channel-fill deposits in ephemeral stream channels. (Schumm, 1960). Note change of vertical exaggeration and characteristic width-depth ratio for each channel type.

It was possible to quantify this relation by using an index of sediment type, the percentage of silt and clay in the perimeter of the channel. This value is both an index of the type of sediment transported through the channels (Schumm, 1977) and an indication of bank stability. Selecting stable reaches of the channels it was possible to establish the following relation between width-depth ratio (F) and sediment types:

$$W/D = 255\ M^{-1.08} \tag{5-1}$$

where M is the percentage of silt and clay in the channel perimeter.

It was noted also that in the channels with high silt-clay loads (Sage Creek) vegetation established itself relatively rapidly in fine sediments deposited on the banks and on the channel floor. The vegetation seemed to grow only in areas where deposition has occurred since channel incision. The vegetation, therefore, not only may aid in further deposition, but is an indication that deposition is occurring. Establishment of vegetation in the less cohesive highly mobile sands was much more difficult and the Arroyo Calabasas' channel was remarkably free of vegetation even when aggradation was occurring. Nevertheless, as soon as a veneer of fine sediment was deposited on the sand, vegetation began to encroach on the channel and eventually completely covered it.

Vegetation, in these channels seems to accelerate the depositional process, but it is not the cause of it. In fact, deposition must be occurring before the vegetation can become established.

Of particular interest here is the behavior of the depositing and eroding channels of each type. As deposition began, the sand was deposited in essentially horizontal layers in the floor of the channel and as this coarse material was deposited the percentage of silt and clay increased in the downstream direction. The increased amount of finer sediment in the channel caused lateral deposition on the channel sides (Fig. 5-8). Two types of deposition continued with fines being deposited on the banks and the coarser sediments on the floor of the channel. During the later stages of filling, all the deposited material was silt and clay, with the coarser sands and gravels being deposited farther upstream. The deposition was clearly by backfilling from a channel plug. Arroyo Calabasas drains an area of coarse sediment and the silt clay in the channel was less than 10% and width-depth ratios were high (80-90). During aggradation the channel appeared to fill from bottom to top (Fig. 5-8).

The nature of the sediment transported through these arroyos and the resulting effect of this material on bank stability significantly affects the morphology of the channel and the nature of the channel filling. In addition, in the early stages of erosion, incision into the fine-grain material produces a roughly rectangular trench of low width-depth ratio, whereas in the sandier alluvium initial incision leads to relatively rapid widening and severe bank erosion. This contrast has been noted between Rio Puerco, which initially was a narrow deep trench and Rio Salado which is a tributary to Rio Grande entering it a short distance to the south of Rio Puerco. When Rio Salado trenched it widened very rapidly. Arroyo Salado drains a very sandy area, and its bank stability is low.

The great difference in channel character and behavior led to further studies that resulted in a classification of stream channels based on the type of sediment transported by the channel (Table 5-2).

Watson and Harvey (1984) reexamined the data of Schumm (1961) and Patton and Schumm (1981) to see whether the morphologies of Sage and Sand Creeks, Arroyo Calabasas and Bayou Gulch could be predicted. Utilizing a technique that involved location for time-for-space substitution (Paine, 1984 in press), they established that certain reaches within each of the channels had reestablished quasi-equilibrium status following incision.

Data available from Schumm (1960) and Patton and Schumm (1981) included a longitudinal profile with cross-sections of each channel, drainage area above each section, D_{50} of the bed material, a weighted percentage of silt-clay in the bed and banks (M), and a classification of the stability of each cross-section.

Henderson (1961) and Wolman and Brush (1961) demonstrated that the wetted perimeter of the channel increased with stream power which, is

Mode of sediment transport and type of channel	Channel sediment (M) (percent)	Bedload (percentage of total load)	Channel stability		
			Stable (graded stream)	Depositing (excess load)	Eroding (deficiency of load)
Suspended load	>20	<3	Stable suspended-load channel. Width/depth ratio <10; sinuosity usually >2.0; gradient, relatively gentle	Depositing suspended load channel. Major deposition on banks cause narrowing of channel; initial streambed deposition minor	Eroding suspended-load channel. Streambed erosion predominant; initial channel widening minor
Mixed load	5–20	3–11	Stable mixed-load channel. Width/depth ratio >10, <40; sinuosity usually <2.0, >1.3; gradient, moderate	Depositing mixed-load channel. Initial major deposition on banks followed by streambed deposition	Eroding mixed-load channel. Initial streambed erosion followed by channel widening
Bed load	<5	>11	Stable bed-load channel. Width/depth ratio >40; sinuosity usually <1.3; gradient, relatively steep	Depositing bed-load channel. Streambed deposition and island formation	Eroding bed-load channel. Little streambed erosion; channel widening predominant

Table 5-2. Classification of Alluvial Channels.
(Table from Schumm, 1977)

defined as a discharge-slope product. Henderson included the median grain size (D_{50}) of the bed material to improve the relationship between wetted perimeter and stream power:

$$\text{Wetted Perimeter} = Q \cdot S^{b} D_{50}^{\ c} \tag{5-2}$$

Richards (1983) stated that channel top width is approximately proportional to wetted perimeter for wide channels.

Hydrologic records are not available for these four channels; therefore, the product of drainage area and channel slope was used as the surrogate for stream power since drainage area is a surrogate for discharge and channel slope is a surrogate for energy slope (Park, 1977). Schumm et al (1981) and Harvey et al (1983) have made this substitution, referring to the drainage area-slope product as the Area Gradient Index (AGI = A·S).

Schumm (1960) recognized that the median grain size (D_{50}) as a single descriptive parameter of the channel sediments failed to account for the sorting of the material. In addition, he suggested that sediment less than 0.074 mm in size greatly influenced the physical behavior of the channel. The result of that study showed that the width/depth ratio (F), which is defined as the bankfull top width divided by the bankfull maximum depth, in stable channels was a function of the weighted mean percentage of silt-clay (M) in the perimeter of the channel:

$$F = 255\ M^{-1.08}$$

Based on this review of literature, it was decided to use the data to predict top width, T.W., from a multi-linear regression of the form:

$$\text{T.W.} = a\ \text{AGI}^{b}\ D_{50}^{\ c}\ M^{d} \tag{5-3}$$

The result of the multi-linear regression is as follows:

$$\text{T.W.} = 53.2\ \text{AGI}^{.26}\ D_{50}^{\ .24}\ M^{-.16} \tag{5-4}$$

The correlation coefficient (r) for the regression is 0.91 indicating a strong relationship. Each of the independent variables also strongly correlated with top width. The data used in this analysis cover a reasonably wide spectrum; widths ranged from 4.9 m to 63.1 m, AGI values ranged from 0.024 km^2 to 0.735 km^2, M values ranged from 2% to 79% and D_{50} values ranged from 0.06 mm to 1.00 mm. As previously stated, all these channels are located in the arid and semi-arid western U.S. and therefore the ability to define new equilibrium top width of the channels following destabilization affords two opportunities. The first, enables a determination to be made with respect to the eventual top width of a currently eroding channel (Equation 5-4). The second, enables a determination of the volume of voided sediment to be made since a relationship between increase in top width and the volume of voided sediment can be generated (Schumm et al, 1984).

CHANNELIZED STREAMS

In many parts of the United States, particularly in the midwest, the early settlers found that the alluvial valley floors were swampy and poorly drained; this was an unhealthy situation and the very fertile bottom lands could not be effectively utilized for agricultural purposes until flooding was eliminated or reduced (Hay and Stall, 1974). Initially, the local people banded together to form drainage districts. For the most part, they straightened the existing sinuous channel by converting it to a ditch. This process is known as channelization. In general, channelizing a river consists of deepening, widening, and straightening of the existing channel. Barnard (1977) refers to basically four types of channel modification methods: 1. Widening, deepening and straightening. This involves a near total physical alteration of stream bed and bank characteristics and channel location. 2. Clearing and snagging which involves removal of brush, logs, and other flow obstructions both in the channel and on the banks. Normally, no channel enlargement is involved, but the increased flow velocity resulting from the reduction of channel roughness increases channel capacity and erosive capability. 3. Diking, where the construction of artificial levees increase the bank height, thereby increasing channel capacity and raising the height of the bankfull stage. 4. Bank stabilization is frequently incorporated with one or more of the other methods. It normally involves the reinforcement of erosion-prone banks with riprap or grass. Channel lining is the ultimate in bank stabilization. The channel is lined with concrete and other material and becomes in effect a "bedrock" conduit.

Most of the above techniques drastically alter the hydraulic characteristics of the channel and channel morphology. Channelization results in a considerable shortening of the channel and a steepening of its gradient. This causes a significant increase in stream power and the erosive capabilities of the flow, which in turn, lead to bank erosion, channel deepening and the attempt of the channel to readjust to the steeper slope (Emerson, 1971). The effects of channelization on both the physical environment and biota of streams are summarized in Figure 5-9.

It should also be noted that channelization is frequently used as a response to altered hydrologic conditions within a watershed. Agricultural activities produce larger quantities of sediment, greater runoff, higher flood peaks and the increased flooding of the bottomlands. The deeper, steeper, straighter channelized channels can accommodate the altered discharge from the watershed. The fact that the channels have deepened indicates a rejuvenation of the watershed. Therefore, channelization further alters the watershed conditions. In addition, if all the water is now confined to a channel, the frequency of overtopping of the banks and inundation of the floodplain is decreased, but the peak discharge within the channel is increased (Campbell et al, 1972). Further incision and modification of the channel will result as the flood peaks increase.

Campbell et al (1972) studied the effects of channelization on the hydrologic character of Boyer River in Western Iowa. Boyer River is the second largest stream in Iowa with a drainage area of 1,188 square miles. Prior to straightening the river's length was about 250 miles in the early 1900's but it is now only 100 miles long. Sinuosity, therefore, has decreased from about 2.5 to 1.0, and the channel gradient has increased 2.5 times. Valley slope ranged from 20 feet per mile in the source area to 1.85 feet per mile near the mouth.

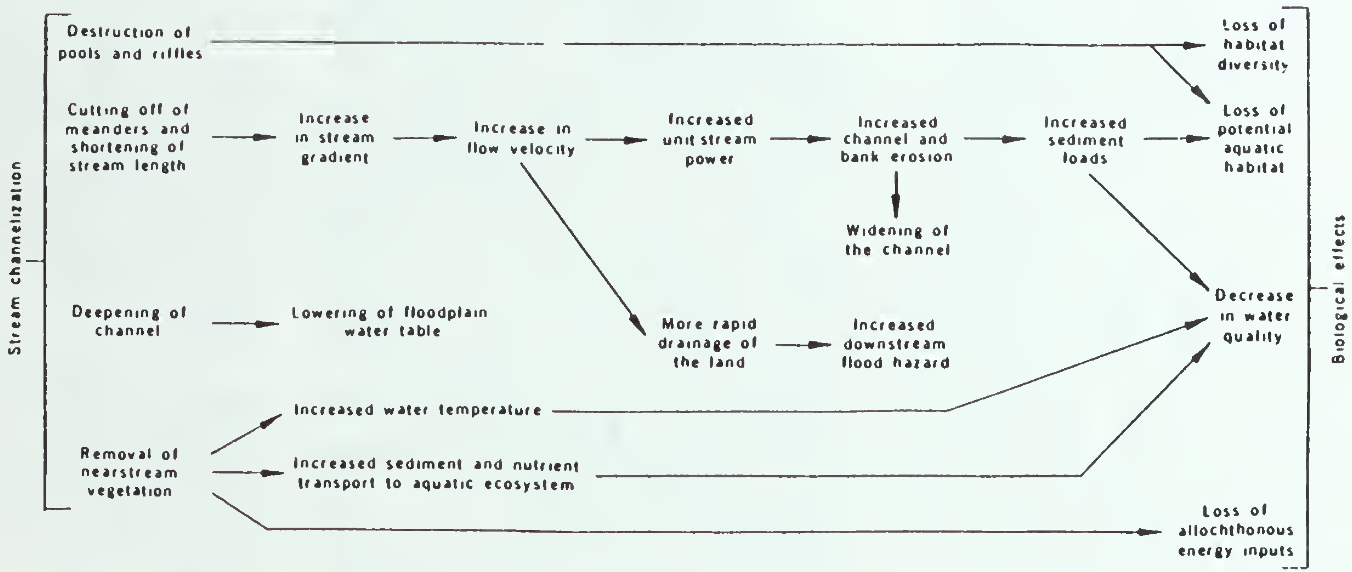

Figure 5-9. Effects of channelization on the physical environment and biota of stream. (Karr and Schlosse, 1978)

To summarize, Campbell et al (1972) conclude that following channelization, peak discharge increased 90 to 190% depending on flood plain roughness. Flooding, which persisted for 30 to 40 hours was eliminated by channelization and the flood-wave travel time was reduced by about 65%.

Behavior of a channelized stream will depend not only on extent of the modification, but also on the character of the bank materials, their erodibility and stratification, as demonstrated for Sage Creek, Sand Creek, and Arroyo Calabasas. In addition, the nature of the sediment load moved through the channel and the change in the character of that sediment load as the drainage basin adjusts will be significant, as indicated by Figure 5-8 and Table 5-2. Initially, the sediment should be derived from channel incision. This will be followed by major bank collapse and production of sediment from the channel sides. Finally, sediment will be derived from upstream rejuvenation of the drainage system or from the upstream agricultural areas.

Adjacent land use will be important. If a border of trees or other deeply rooted vegetation remains along the banks of the channel it will be considerably more stable than if the vegetation is removed and the former floodplain is farmed to the bank of the channel.

Depending on the extent of the incision and bank erosion, the channel may simply incise a few feet and stabilize. At the other extreme, the depth of incision may be so great and the extent of bank erosion so significant that a completely new cycle of erosion is instituted (Rio Puerco). In this case, the end result will be creation of a new floodplain at a lower level.

Examples of drastic channel incision and enlargement as a result of channelization are provided by Emerson (1971; Blackwater River), Piest et al (1977; Tarkio River), Daniels (1960; Willow Creek; 1966, Thompson Creek), Barnard (1977; Big Pine Creek) and Harvey et al (1983; Oaklimiter Creek). Tarkio River flows southwest through southwestern Iowa and northwestern Missouri to its confluence with the Missouri River. It has a drainage area about 1400 km (540 sq. mi.), that is underlain by thick loess deposits.

The confluence of Tarkio River with Missouri River has been drastically changed, and its confluence is now 25.7 km upstream from its former location. This was accomplished by cutting a channel directly from the mouth of the drainage basin across the floodplain of Missouri River. Drainage districts were formed before 1900 and channel straightening operations by individual farmers or small groups occurred even earlier. Most of the lower and middle channel reaches of Tarkio River were straightened by 1920. As a result, Tarkio Creek, its tributaries and little Tarkio Creek have been subjected to significant changes of channel width and depth (Fig. 5-10).

The changes in the Tarkio drainage system are what can be expected as a result of the steepening of the gradient of the channel both by upstream relocation of the mouth of the channel and straightening of the river itself. However, it is likely that a degree of stability has been achieved by the channels as they widened from about 10 m to 50 m and as the banks of the channel decreased in steepness (Fig. 5-10).

WILLOW CREEK

Willow Creek, Iowa is a tributary to Boyer River described previously, and it is north of Tarkio River (Daniels, 1960). Thompson Creek is a tributary to Willow Creek. In 1853 Willow Creek meandered across the valley in a six to seven-feet wide channel, about six feet below the floodplain.

Willow River frequently flooded its valley to a depth of several feet and the valley floor was considered unfit for cultivation. To stop the flooding, a drainage ditch was dug to replace the meandering channel. Construction began in 1916 and the upper part of the ditch was completed in 1920. The ditch in the Willow River Valley began to deepen and widen soon after construction. The 1966 channel, is not like the original channel. The ditch was built with a flat bottom with banks sloping 45^{o}. The 1966 channel is U-shaped, has numerous silt, sand and gravel bars below water level and a few small meanders have formed.

Where the banks slope less than 45^{o} they are almost completely covered by grass, trees and weeds; the vertical banks are bare. The

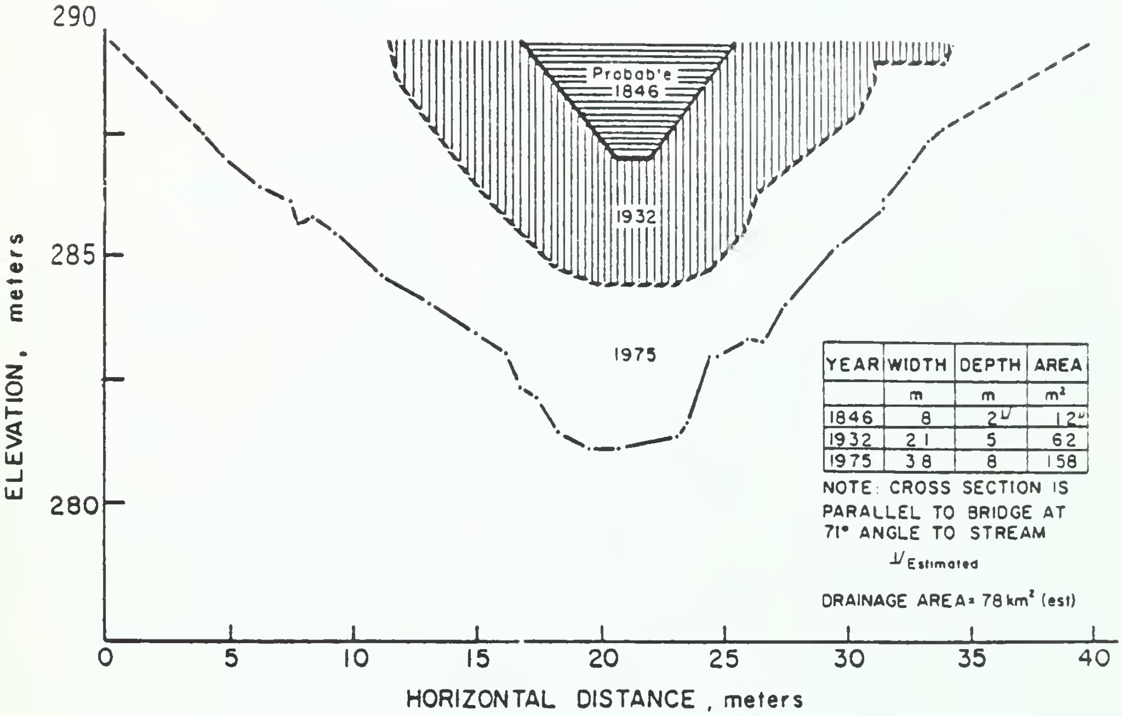

Figure 5-10. Cross section showing modern trenching of an Atchison Co., Mo. drainageway. Profile and cross section changes West Tarkio Creek near Missouri-Iowa boundary. (Piest, et al, 1977)

channel floor has a maximum depth of 45 ft. below the floodplain in the central reaches of the stream. The U-shaped cross section of the channel in 1966 was different from that of the Blackwater (Emerson, 1971) and Tarkio channels, and it may reflect a greater cohesiveness of loess in this area. Residents of the Willow River Valley remember a series of nickpoints moving upstream in the drainage ditch. Some of the nickpoints were 10 feet high, and many plunge pools were 10 feet deep. These nickpoints were most numerous in the lower part of Willow Valley in the early 1930s.

Since 1940 the Willow River drainage ditch has been relatively stable from mile 36 to 26 (Fig. 5-11). However, upstream the gradient of the ditch is considerably steeper, and Daniels predicted that additional degradation of up to 40 feet would occur in the upper reaches of Willow River (Fig. 5-12).

Observations made in 1980 in the lower reaches of Willow River indicate that a new floodplain has formed at the bottom of the trench, and vegetation has stabilized the banks. The Willow drainage ditch may have achieved some degree of stability because the slope of the easily erodible loess banks have declined, and they have become vegetated.

Daniels' study of the erosional history of this area led him to the conclusion that there is a general similarity between past and present cycles of cutting and filling in the Willow River valley. Based on these studies Daniels (1966, p. 81) makes the following astute statement: "I suggest that control measures designed to reduce the effects of stream trenching in Thompson Creek may be of little long-range value if the stream is at or near grade. But, they may be of considerable value in reducing the trenching predicted in the Willow River Valley upstream from the Willow drainage ditch. It seems reasonable to concentrate control measures in areas of potential trenching rather than areas that may be nearly through the trenching cycle." Daniels' approach is geomorphic in nature, and he provides an understanding of the overall problem of stream entrenchment and gully development in the loess areas of Iowa.

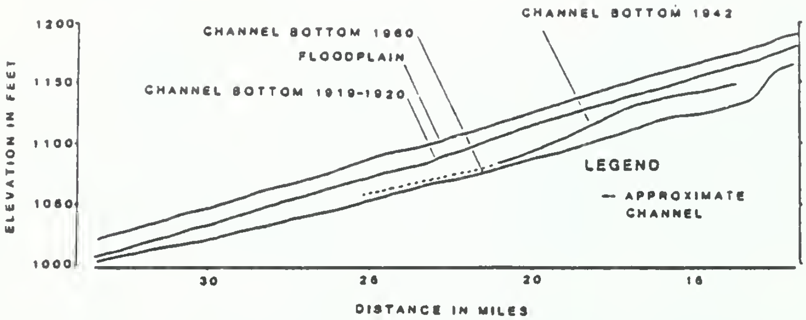

Figure 5-11. Profile of Willow drainage ditch in Willow River valley and post-construction changes. (Daniels, 1966)

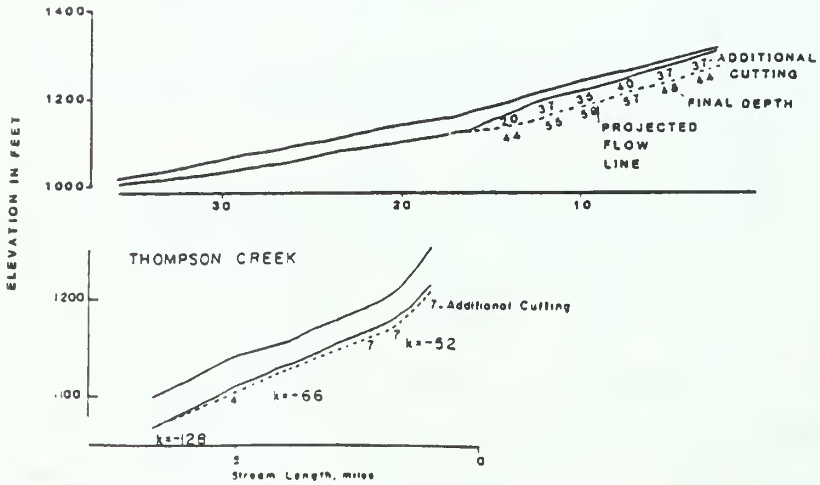

Figure 5-12. Predicted profiles of the Willow River and drainage ditch and Thompson Creek. (Daniels, 1966)

BIG PINE CREEK INDIANA

A detailed study of the response of a channel to straightening is provided by Barnard (1977) for Big Pine Creek and Big Pine Ditch in Benton County, Indiana. Big Pine Creek flows within the Tipton Till Plain, a flat Wisconsin-age plain composed of glacial till and isolated areas of sand and gravel outwash deposits. The surface materials and soils are derived from glacial drift, water-deposited outwash and recent alluvial and eolean deposits. Depth to bedrock varies considerably across the county as the pre-glacial topography was more rugged than it is today.

Big Pine Creek drainage basin has an area of 15.9 sq. mi.; basin relief is 115 feet and drainage density is 3.56. Relatively little information is available on the natural morphology of the stream before 1932. However, Pine Creek prior to straightening was a small stream with a sinuosity of 1.42. Meander wavelength varied considerably, averaging 250 feet. The average slope was 6.3 ft. per mile or 0.00119. Bedrock is exposed in the channel about midway along the longitudinal profile (Fig. 5-13). In 1930 it was proposed that the Big Pine Creek be channelized. All trees and brush were to be cleared to allow 125 feet on right-of-way and spoil piles were to be evenly distributed on both banks, leaving a 5-foot berm between the edge of ditch banks and the toe of the spoil banks.

Channelization reduced total mainstream length from 10.1 to 7.2 miles or by 30%, so average gradients were increased from 6.3 to 7.9 ft/mile or from .00119 to .00161. Construction of the ditch was completed by June 1932. In the six years immediately following completion of the project the channel adjusted drastically, especially where gradients where high (Fig. 5-13). The bedrock control stabilized the upstream reach. In this particular case, the channel was adjusting in steeper reaches to a drastically increased gradient, and this adjustment took place both by incision, widening and development of higher sinuosity up to 1.2 in one of the more rapidly eroding reaches. In this reach a new floodplain and vegetated point bars have developed (Fig. 5-14). A cross section close to the headwaters shows only minor

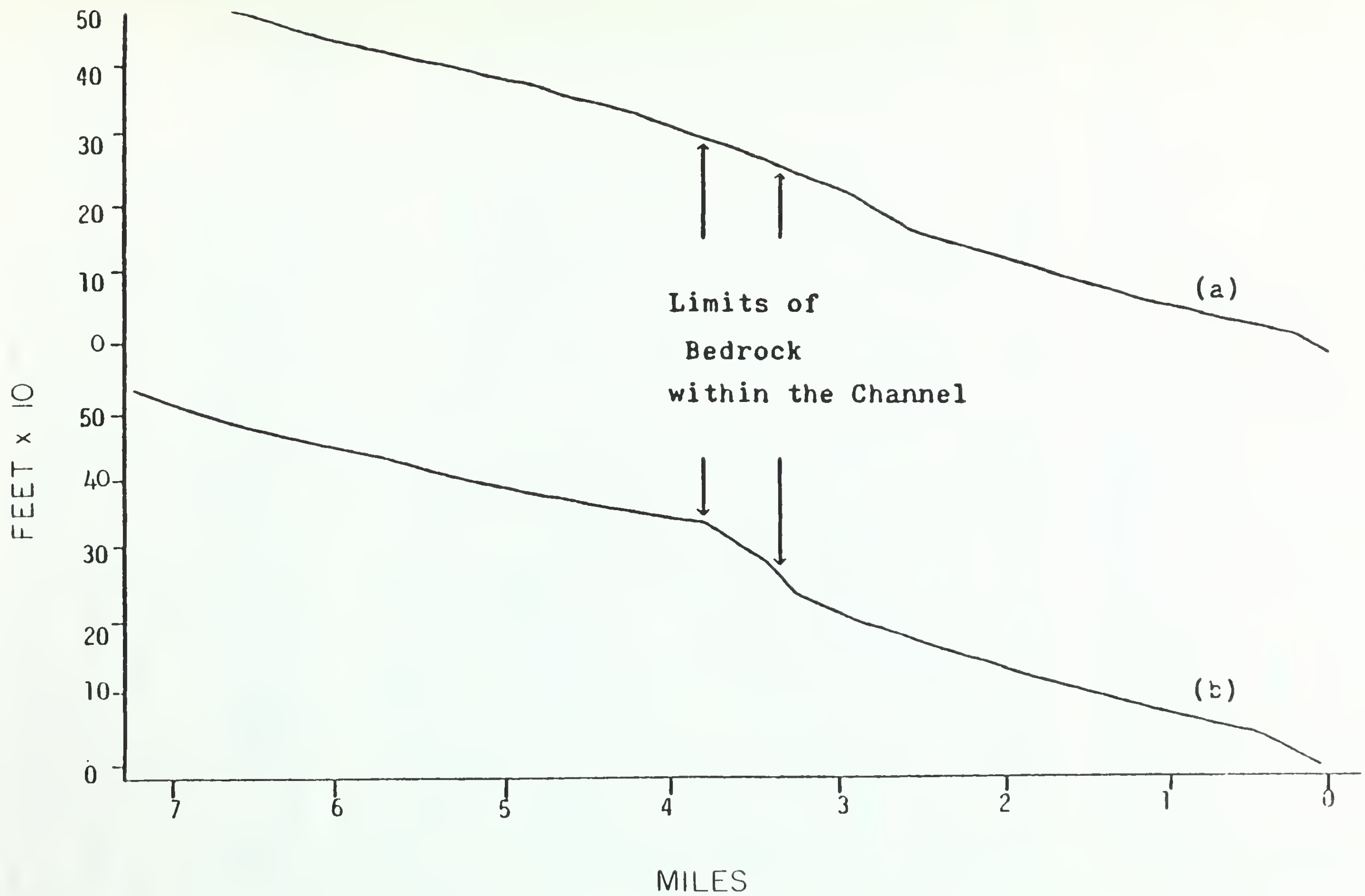

Figure 5-13. Channel profile of Big Pine Creek Ditch: a) after channelization in 1932 (Benton Co. Ditch Docket #137); b) after 30 years of recovery in 1962. (Barnard, 1977)

changes since 1932 (Fig. 5-14a) but downstream sections show considerable bank slumping and deposition on the inside of meander bends (Fig. 5-14b). Only a short distance upstream from the end of the ditch where gradients are steep, considerable enlargement of the channel has occurred with development of a floodplain (Fig. 5-14C).

In summary, the Big Pine Creek ditch seems to have behaved as have other channelized streams. The details of the adjustment vary as a result of the differences in the material into which the channels are incising.

Barnard proceeds beyond the efforts of other investigators in order to explain variation in the response of Big Pine Creek to channelization. In order to do this he utilizes the concept of stream pattern thresholds (Schumm, 1977) and the relation between valley slope, or stream power and sinuosity (Fig. 5-15). When his data for sinuosity are plotted against valley-floor gradient (Fig. 5-16), sinuosity decreases with decreasing valley gradient. This is true also for stream power (Barnard, 1977; Barnard and Melhorn, 1982). The range of values is not sufficient to define the thresholds between straight meandering and braided streams (Fig. 5-15), but the range of sinuosity reflects valley slope variability and the maintenance of a relatively constant stream gradient by sinuosity adjustments. This suggests that channel adjustment after channelization will vary greatly depending on valley slope and sinuosity of a reach. Sinuosity changed between the time of channelization and 1938, 1963 and 1971. The steeper reaches adjusted to a greater extent and more rapidly (Fig. 5-16).

According to Barnard (1977, p. 67) if Big Pine Creek was in a nearequilibrium state prior to channelization, the envelope developed for that time period represents the stream in its most stable condition (Fig. 5-16). Deviations from this envelope represent instability and thus indicate that readjustment is forthcoming. Channelization produces an extreme deviation by reducing sinuosity to one (Fig. 5-15).

Barnard (1977) fitted lines by visual inspection to each set of points representing the various time periods (Fig. 5-17). The change of sinuosity in the direction of the prechannelized state is evident in the lines that represent sinuosity in 1938, 1963, and 1971.

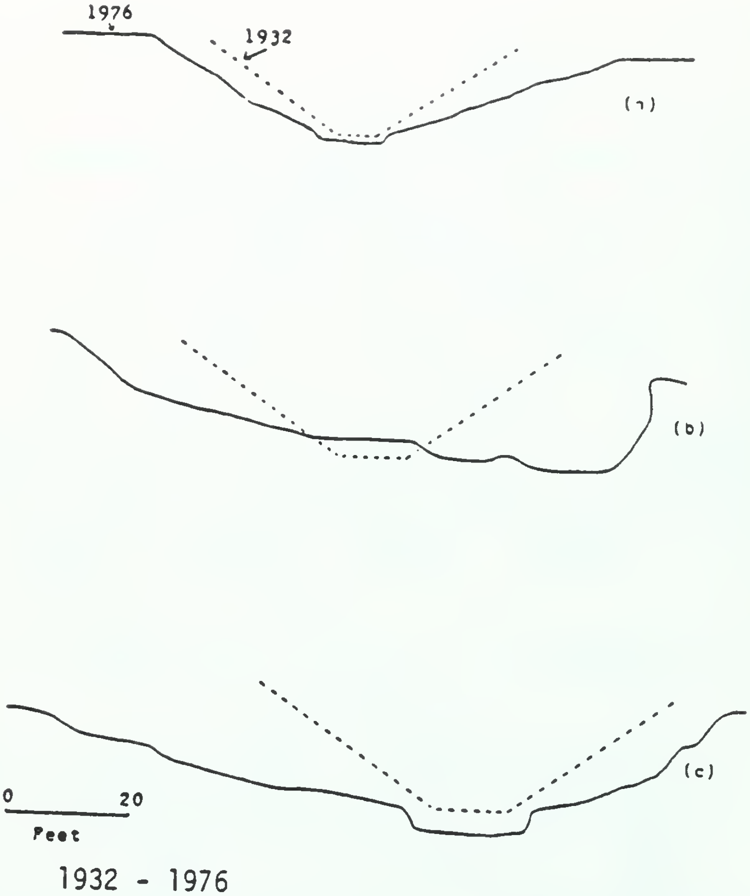

Figure 5-14. Cross sections in: a) Section 24, T26N, R8W, b) Section 34, T26N, R7W (downstream end); c) Section 1, T25N, 0.5 mi. upstream from outlet. (These cross sections are approximations) (Barnard, 1977)

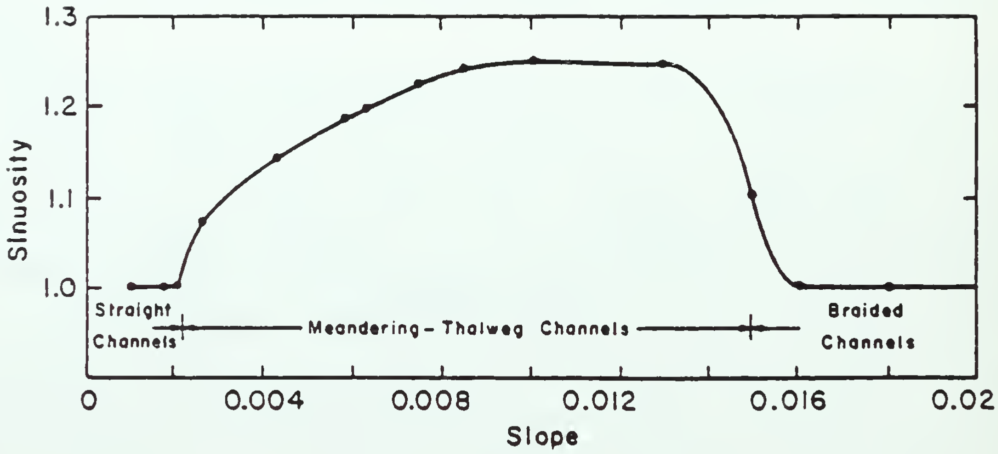

Figure 5-15. Relation between flume slope and sinuosity (ratio of channel length to length of flume or length of valley) during experiments. The change from a straight to a sinuous pattern and from a braided pattern occurs at two threshold slopes. The absolute value of slope at which such changes occur will be influenced by discharge. Discharge was maintained at 0.15 cfs during the experiments. (Schumm and Khan, 1972)

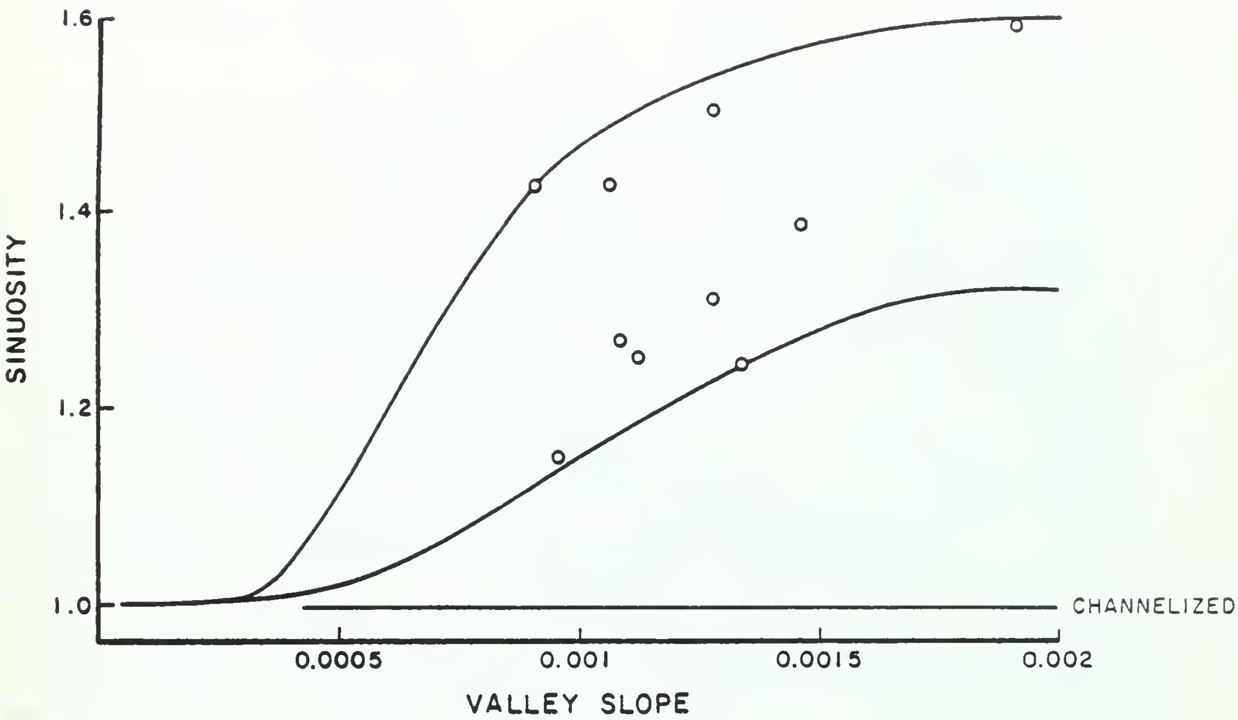

Figure 5-16. Variations of sinuosity with slope for pre-channelized conditions on Big Pine Creek Ditch. (Barnard, 1977)

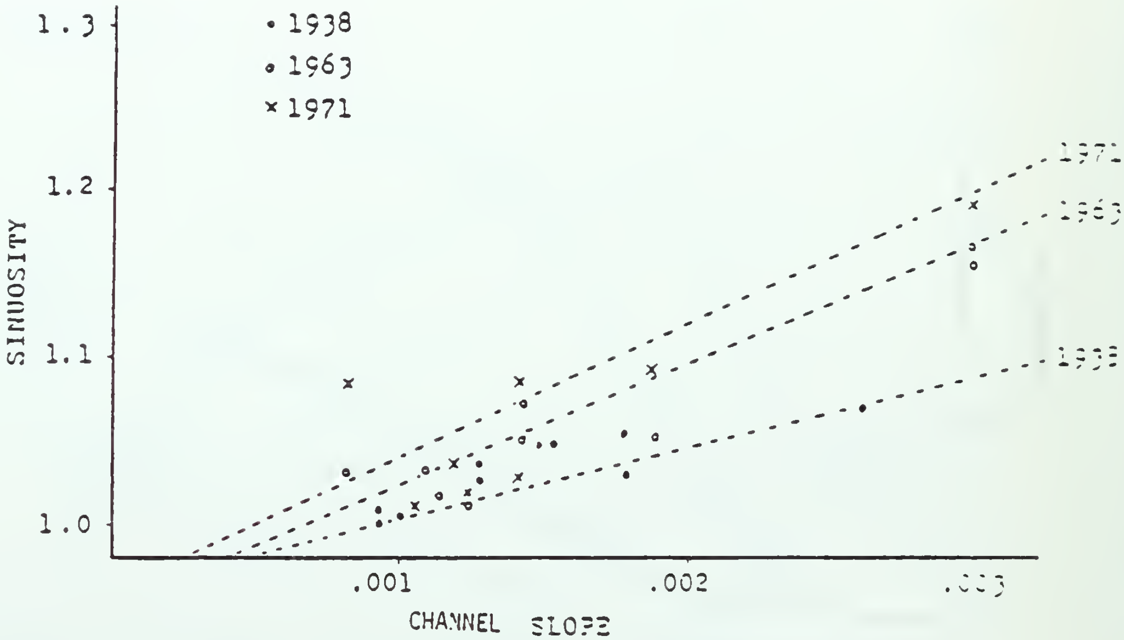

Figure 5-17. Variations of sinuosity with slope in Big Pine Creek Ditch in 1938, 1963 and 1971. (Barnard, 1977)

Barnard (1977) calculated a rate of recovery and he concludes that by the year 2080 or after 165 years of recovery Big Pine Ditch, if allowed to evolve without interruption should again be in a new equilibrium state.

OAKLIMITER CREEK, MISSISSIPPI

Schumm et al (1981, 1984) and Harvey et al (1983) report on the 15-year response of Oaklimiter Creek to channelization that occurred in 1965. Oaklimiter Creek is located in Benton County, northern Mississippi and it has a drainage area of 42 square miles. It lies within the North-Central Hills subdivision of the Gulf Coastal Plain physiographic province (Fenneman, 1932). Oaklimiter Creek was channelized to reduce the frequency of out of bank flows which were reported to occur about 10 times per year. Prior to 1965 there had been very little modification of the natural channel which had sinuosities ranging from 1.3 to 2.5.

Table 5-3 summarizes the response of Oaklimiter Creek to the channelization after 15 years (1980). In general channel slopes had reduced from the 1965 values (columns 2 and 3). The reduction in slope was accomplished by bed degradation which resulted in the development of an incised channel. The degree of incision can be determined by reference to columns 11 and 12. The increased channel depths caused bank failures, which resulted in channel widening, the magnitude of which can be seen in columns 8 and 9. The combination of increased depth and width resulted in significant increases in cross-section area (columns 14 and 15). The relative magnitudes in the changes of channel slopes, widths, depths and cross-section areas can be appreciated when the 1980 and 1965 values are ratioed (Table 5-4).

An evolutionary sequence of channel cross sections which were developed by the use of a location-for-time substitution technique (Paine, 1984 in press) was formulated into a channel evolution model (Fig. 5-18) that yielded a sequence of five channel reach types (Types I - V). The reach types are associated with conditions ranging from total disequilibrium to a state of quasi-equilibrium. Figure 5-18

Station in feet	Channel Slope			Bottom Width in feet			Top Width in feet			Depth in feet			Cross Section Area in square feet			Width-Depth Ratio			Drainage Area in sq. miles
	1965	1980		1965	1980		1965	1980		1965	1980		1965	1980		1965	1980		
		$\bar{x}$	s		$\bar{x}$	s		$\bar{x}$	s		$\bar{x}$	s		$\bar{x}$	s		$\bar{x}$	s	
(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)	(12)	(13)	(14)	(15)	(16)	(17)	(18)	(19)	(20)
0+00 to 97+00	0.0015	0.0009	0.0002	36	88	15	52	128	15	8.0	12.5	1.2	352	1346	200	6.5	10.3	1.6	41.34
97+00 to 202+00	0.0016	0.0008	0.0002	36	76	13	52	140	18	8.0	19.5	1.9	352	2119	412	6.5	7.2	0.7	38.23
202+00 to 299+00	0.0017	0.0006	0.0003	34	72	10	48	131	20	7.0	15.6	1.8	287	1583	315	6.9	8.5	1.7	33.75
299+00 to 439+00	0.0018	0.0010	0.0002	24	61	14	42	108	20	7.0	16.5	2.9	231	1520	496	6.0	6.6	1.3	22.42
439+00 to 492+00	0.0019	0.0015	0.0004	24	52	17	37	108	21	6.5	20.1	2.4	198	1634	535	5.7	5.4	1.0	17.94
492+00 to 518+00	0.0019	0.0067	0.0012	22	47	17	34	88	2	6.0	21.9	1.7	168	1471	73	5.7	4.0	0.3	15.03
518+00 to 574+00	0.0019	0.0023	0.0019	20	51	17	32	104	32	6.0	18.6	2.8	156	1453	478	5.3	5.5	1.4	12.41
574+00 to 585+00	0.0019	0.0013	0.0002	20	54	9	32	97	25	6.0	17.8	1.1	156	1432	84	5.3	5.4	1.8	10.49
585+00 to 611+00	0.0025	0.0012	0.0001	20	36	3	32	90	22	6.0	18.4	2.1	156	1154	246	5.3	5.0	1.4	10.33
611+00 to 626+00	0.0025	0.0015	0.0007	20	44	8	30	93	16	5.0	17.8	1.4	125	1211	155	6.0	5.1	0.9	9.93
626+00 to 657+00	0.0025	0.0026	0.0016	20	26	4	29	63	9	4.5	15.4	3.7	110	689	262	6.4	3.7	0.2	8.28
657+00 to 676+00	0.0025	0.0011	0.0008	20	35	5	28	63	8	4.0	15.9	1.8	96	832	129	7.0	3.9	0.1	6.98
676+00 to 763+00	0.0025	0.0012	0.0006	10	29	10	20	58	16	5.0	13.5	3.3	75	590	222	4.0	4.5	1.7	3.78
763+00 to 781+00	0.0025	0.0017	0.0002	10	24	5	20	39	8	5.0	9.2	1.7	75	294	103	4.0	4.3	0.2	2.52
781+00 to 829+00	0.0030	0.0016	0.0003	8	20	4	16	36	7	4.0	9.7	0.4	48	274	61	4.0	3.7	0.6	1.90

NOTE: 1 ft = 0.305 m; 1 sq ft = 0.093 m^2; 1 sq mi = 2.59 km^2

Table 5-3. Summary of morphometric variables for Oaklimiter Creek, in 1965 and 1980. The data is segregated on the basis of identifiably different construction reaches. 1980 data are means ($\bar{x}$) and standard deviations (s) derived from cross-section surveys located in the construction reaches.

Station in feet (1)	Ratio of Channel Slopes 1980/1965 (2)	Ratio of Bottom Widths, 1980/1965 (3)	Ratio of Top Widths 1980/1965 (4)	Ratio of Depths 1980/1965 (5)	Ratio of Cross Section Areas 1980/1965 (6)
0+00 to 97+00	0.6	2.4	2.5	1.6	3.8
97+00 to 202+00	0.5	2.1	2.7	2.4	6.0
202+00 to 299+00	0.4	2.1	2.7	2.2	5.5
299+00 to 429+00	0.6	2.5	2.6	2.4	6.6
439+00 to 492+00	0.8	2.2	2.9	3.1	8.3
492+00 to 518+00	3.5	2.1	2.6	3.7	8.8
518+00 to 574+00	1.2	2.6	3.3	3.1	9.3
574+00 to 585+00	0.7	2.7	3.0	3.0	9.2
585+00 to 611+00	0.5	1.8	2.8	3.1	7.4
611+00 to 626+00	0.6	2.2	3.1	3.6	9.7
626+00 to 657+00	1.1	1.3	2.2	3.4	6.3
657+00 to 676+00	0.4	1.8	2.3	4.0	8.7
676+00 to 763+00	0.5	2.9	2.9	2.7	7.9
763+00 to 781+00	0.7	2.4	2.0	1.8	3.9
781+00 to 829+00	0.5	2.5	2.3	2.4	5.7

NOTE: 1 ft = 0.305 m

Table 5-4. Comparison of channel, slopes, bottom widths, top widths, depths and cross-sectional areas, 1980 and 1965, for the different construction reaches of Oaklimiter Creek. 1980 values are means.

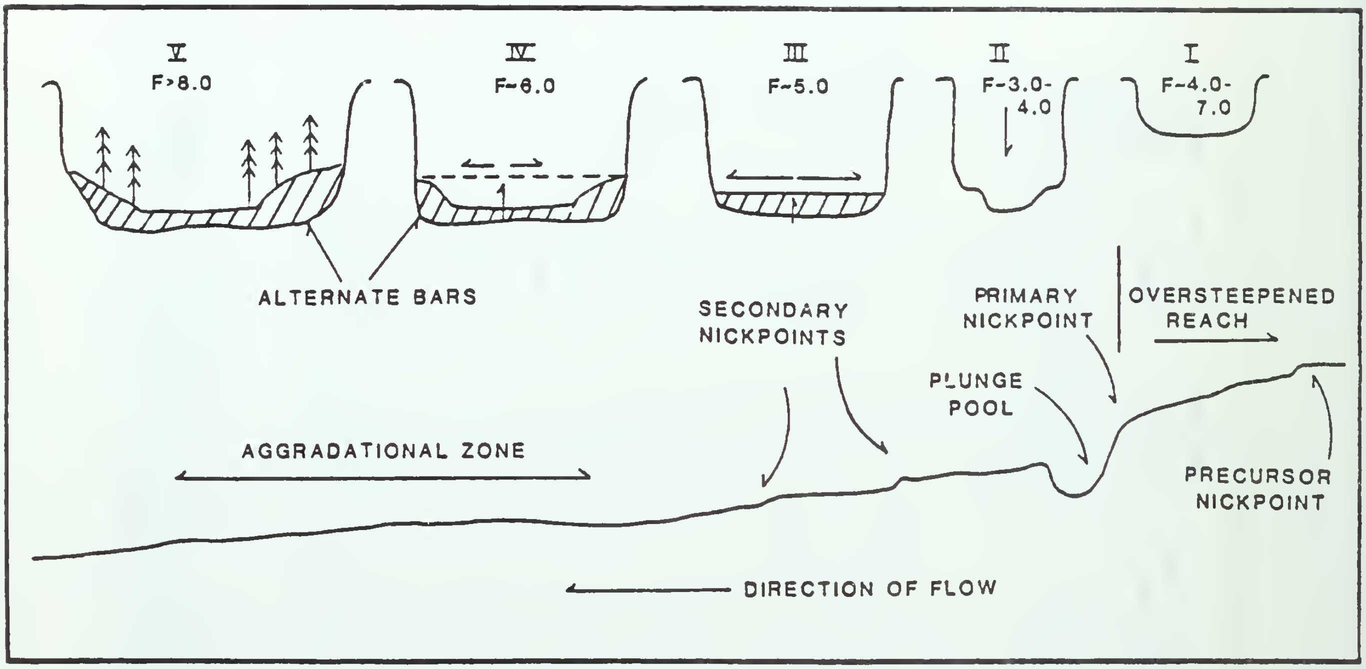

Figure 5-18. Schematic longitudinal profile of an active channel showing identifiable features. Schematic cross section profiles corresponding to reaches on the longitudinal profile show the evolution of the reaches from Type I to Type V. Typical F values are shown. Size of the arrows indicate the relative importance and direction of the dominant processes, degradation, aggradation and lateral bank erosion.

schematically depicts the evolution of the channel reaches through time. Table 5-5 provides a summary of the changing magnitude of the variables and the dominant processes, through time. In Oaklimiter Creek a state approaching quasi-equilibrium was recognized (i.e. Type IV reach) when, (1) the width-depth ratio (F) was about 6, (2) channel slope had reduced to approximately 0.001, (3) sediment accumulation in the bed of the channel had reached about 3 feet, and (4) a meandering thalweg had developed under low-flow conditions.

Oaklimiter Creek therefore appears to have responded to channelization in the same fashion as the other channelized streams. Schumm et al (1981) and Harvey et al (1983) carried their efforts beyond those of the other investigators in that they attempted to use the channel response to date to quantitatively predict the future development of the non-equilibrium reaches. Their predictive relationships were based on two assumptions: (1) that base level for Oaklimiter Creek would remain at its current elevation and (2) that land use in the watershed would not significantly change.

Utilizing only those reaches which had attained a width-depth ratio (F) of 6 or greater, Schumm et al (1981) developed a relationship between the channel depth and the increase in channel top width since construction (Fig. 5-19). Watson and Harvey (in prep.) have demonstrated that an equilibrium top width (T.W.) in the quasi-equilibrium reaches can be predicted by the maximum depth of the channel (d_m) and the change in channel depth since construction (Δd) as follows:

$$T.W. = 1.43\ d_m^{\ 1.96}\ \Delta d^{0.75} \qquad (5\text{-}5).$$

The coefficient of determination (r^2) for the relationship is 0.82. As will be shown later, this relationship has its basis in soil mechanics, but it also suggests that the primary determinant of width in incised channels is the depth of the channel. Because depth is an expression of the degree of incision that has occurred and this is in turn related to channel slope adjustment it is reasonable to conclude that Equation 5-5 can be used to predict incised channel adjustment. In essence the adjustments involve a mutual accommodation of cross-section area and

STAGE	LOCATION	TOP WIDTH in feet	DEPTH in feet	WIDTH/DEPTH RATIO	THALWEG SLOPE	DEPTH OF SEDIMENT in feet	DOMINANT PROCESS
(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)
I	Upstream of nickpoint (580+00)	82	17.3	4.7	0.0020	0	Transport Sediment
II	Immediately downstream of nickpoint (560+00)	82	21.6	3.8	0.0018	variable	Degradation
III	Downstream of II (520+00)	100	20.1	4.9	0.0018	1.5	Rapid Widening
IV	Downstream of of III (450+00)	115	19.2	6.0	0.0016	2.5	Aggradation and Development of Meandering Thalweg
V	Downstream of IV (435+00)	119	15.3	7.8	0.0010	6.3	Aggradation and Stabilization of Alternate Bars

NOTE: 1 ft = 0.305 m

Table 5-5. Summary of morphometric data used to determine channel evolution model, Oaklimiter Creek (Stations 600+00 to 400+00).

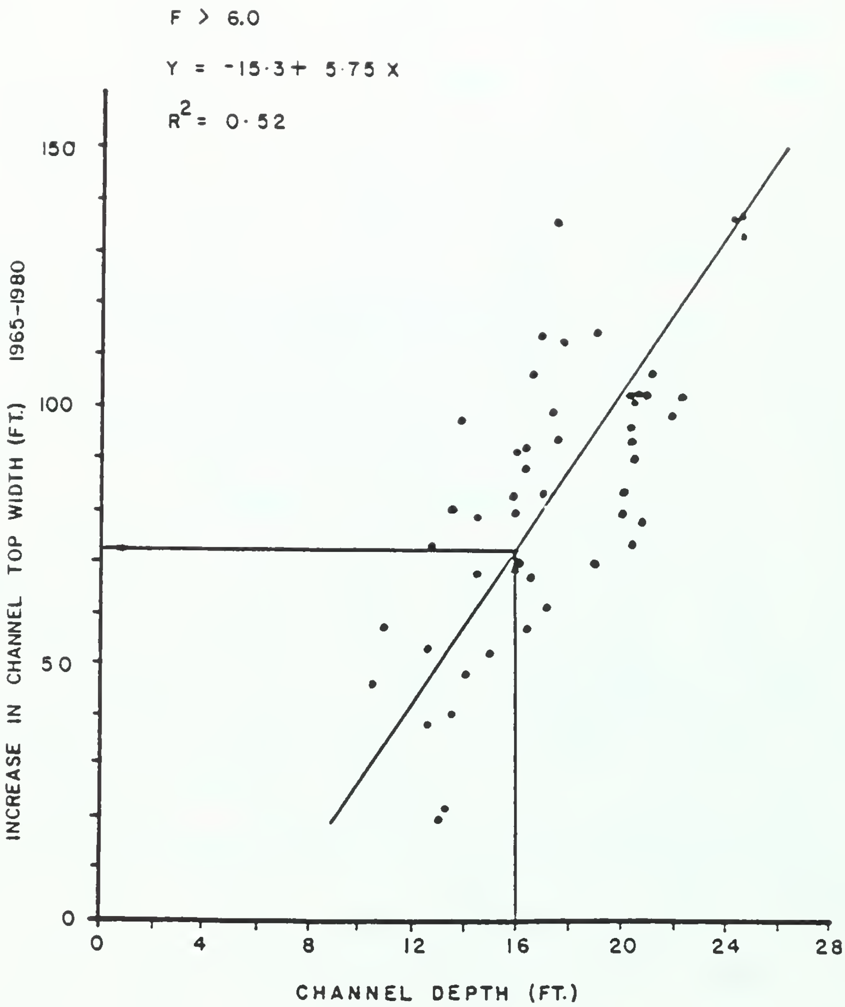

Figure 5-19. Oaklimiter Creek. Increase of channel top width (1965-1980) plotted against channel depth (1980), for F values > 6. Arrows show how to use the figure for predictive purposes.

channel slope to maintain continuity of water and sediment transport (Leopold et al, 1964).

Hydrologic and hydraulic data were not available in Oaklimiter Creek and therefore an energy parameter, Area Gradient Index (AGI) was developed. AGI is the product of the drainage area above a cross section and the channel slope through the cross section. It is a measure of stream power since drainage area is correlated with discharge (Park, 1977) and bed slope provides an estimate of the energy slope (Simons and Senturk, 1977) and the discharge-slope product has been defined as a measure of total stream power (Henderson, 1961; Wolman and Brush, 1961; Richards, 1983).

Channel top widths for both non-equilibrium and quasi-equilibrium reaches in 1980 are plotted against AGI (Fig. 5-20). Upper and lower parallel lines define an envelope of maximum and minimum values and it might be assumed that the quasi-equilibrium top widths are represented by the upper line. However, Schumm et al (1981) argued that this approach would tend to over estimate the channel width. They argued that since by observation channel slope is reduced, as the channel evolves towards a condition of quasi-equilibrium, then since drainage area remains fixed, the AGI value must decrease with time. For example, if the drainage area at a cross section is 10 sq. mi. and the non-equilibrium slope was 0.0025, the the AGI value is 0.025 which represents a channel top width of 75 feet from the lower limiting line (Fig. 5-20). The maximum channel width might be 200 feet, but evolution of the channel reduces AGI to say 0.009, which represents a channel width of 100 feet. This suggests that a further widening of 25 feet will occur before quasi-equilibrium is attained at this reach.

Harvey et al (1983) derived a relationship between channel top width for only quasi-equilibrium reaches and AGI:

$$T.W. = 194 \cdot 98\ (AGI)^{0.12} \qquad (5\text{-}6).$$

The correlation coefficient for this relationship is 0.75. They suggested that since this relationship was derived for quasi-equilibrium reaches then the quasi-equilibrium channel width in currently non-

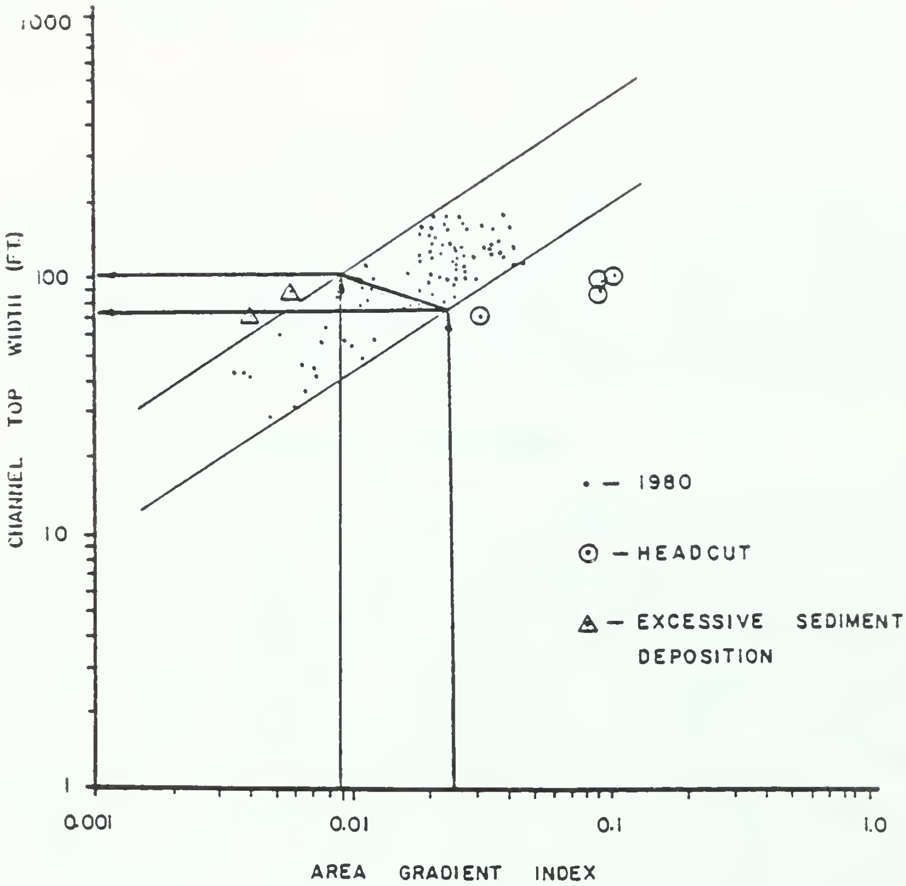

Figure 5-20. Oaklimiter Creek. Channel top width plotted against Area Gradient Index. Upper and lower parallel lines define an envelope of maximum and minimum values, as measured in 1980. Arrows show how to use the figure for predictive purposes.

equilibrium reaches could be predicted from the relationship. This technique of spatial interpolation had been previously used by Park (1977).

Schumm et al (1981) derived a relationship between increase in channel top width (X) and the volume of sediment eroded from the channel (Y) as it evolved:

$$Y = 1.73 + 0.67X \tag{5-7}$$

The coefficient of determination for the relationship is 0.85. Therefore, by determining the expected increase in channel top width (Equations 5-5, 5-6) an estimate of the volume of sediment that will be eroded from the bed and banks of the channel can be made (Equation 5-7).

EXPERIMENTAL STUDIES

In spite of the amount of work that has gone into the study of gullies and entrenched streams in the field, the investigations are hampered because channel changes are related to the frequency and magnitude of hydrologic events, and it is not always possible to be on the scene to determine the effects of these events on a channel. Therefore, the details of the evolution of the incised channel, which are needed before one can institute measures to control channel activity, are not available. Hence, the need for experimental studies.

Several experimental investigations have been carried out in flumes and in the REF at Colorado State University. One experiment (Gardner, 1973) demonstrated that, as expected, channel incision will follow the course of the existing channel. That is, with base-level lowering incision will progress up a meandering channel, thereby developing incised meanders.

In the process of incision Gardner (1973) noted that the meander bends may be deformed as the downstream meander shift is stopped by incision into bedrock, whereas the upstream limb is still shifting downstream in alluvium (Fig. 5-21). Obviously, this process can lead to meander cutoffs and to a decrease of channel sinuosity. The channel pattern may change significantly during incision if alluvial thicknesses

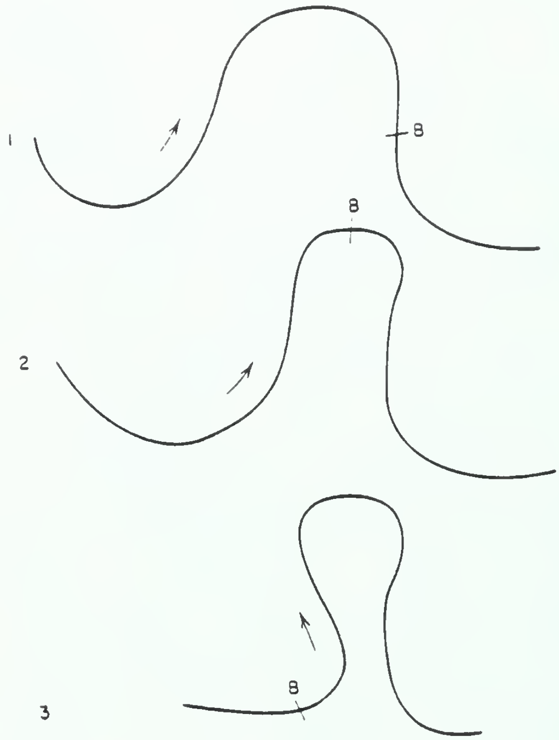

Figure 5-21. Compression of a meander as the channel encounters bed rock (B) which fixes the position of the downstream limb. (Gardner, 1973)

are considerable, however, when the channel incises into bedrock its pattern becomes fixed.

Additional experiments were carried out to determine the effects of base-level lowering on channel morphology (Begin, 1979; Meyer, 1980; Begin et al, 1980 and 1981). In the first experiment a 13 m long channel was fashioned along one side of the REF and water and sediment was delivered to this channel by the application of precipitation (Begin, 1979). After preliminary runs to permit the system to adjust, base level was lowered, and not only was the progression of erosion and channel evolution documented, but changes in sediment yields were monitored. Changes of longitudinal profile during a total of 14 hours of run time were plotted (Fig. 5-22). To initiate the run base level was lowered by 35 mm. This created a vertical headcut that migrated upstream 300 cm after 30 minutes. At a distance of 5 m from the outlet the nickpoint was degrading. Note on Figure 5-22 the steeper reach on the initial profile (t=0) above 5 m. The erosion of this steeper reach is analogous to erosion of critical reaches where discontinuous gullies form in an alluvial valley. The primary nickpoint merged with this reach of discontinuous erosion, and it disappeared. At the end of this run, the channel was steeper but as irregular as the initial profile.

The next run was begun after 120 minutes with an additional 76 mm of base-level lowering. A vertical headcut was formed, and it migrated upstream. After 45 minutes it reached 7.5 m, where it also lost its identity. Above this point degradation occurred, as a series of small nickpoints migrated upstream between 10 and 13 m.

The profiles (Fig. 5-22) show downstream steepening as a result of the two lowerings of base level, and they show that, in general, degradation is a maximum where a profile convexity exists. For example, compare the profile at 240 and 360 minutes and the profile at 660 to 840 minutes. Note aggradation at the downstream end of the 660-minute profile and the four convexities which may be the result of local sediment storage. In the final 840 minutes these convexities have not been removed completely, and the final profile is still irregular, although three hours of run time did not cause appreciable downcutting.

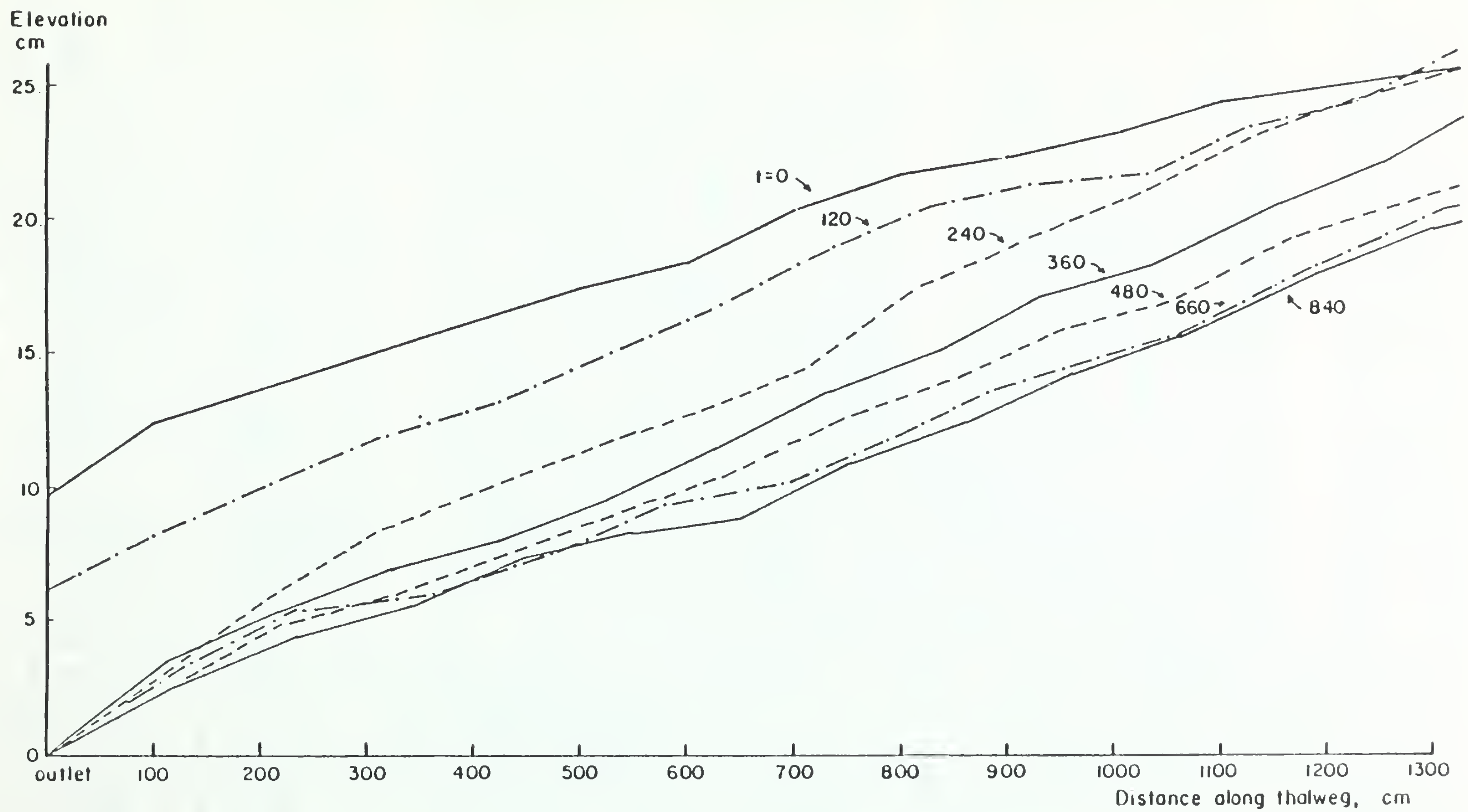

Figure 5-22. Profile development during runs 3A and 4; time t = 0 is 42.00 hrs. Base level was lowered twice: at t = 0 and at t = 120 minutes. Numbers denote time in minutes. Profiles are based on mean bed elevations. (Begin, 1979)

The incision of the experimental channel was generally predictable; however, the development of additional irregularities and convexities was not, and they unquestionably are significant in determining channel behavior during incision.

Another striking aspect of the profiles is that the lower 2 m of the profile remained steep, as compared to upstream reaches. Presumably at some later stage upstream sediment yield will be reduced to the extent that renewed incision will remove the lowermost convexity.

Because base level was lowered twice during these runs it was difficult to determine what caused the nickpoints in the upper part of the channel. After about 250 minutes the vertical headcut was replaced by a nickzone (Fig. 1-1), which apparently caused a reduction in upstream migration of the nickpoint (Fig. 5-23).

Another feature of interest is the secondary nickpoints that developed downstream, as a result of sediment storage in the channel. At 300 minutes three secondary nickpoints were present in the channel. These features add considerable complexity to the channel evolution, but their development is undoubtedly the result of the development of critical reaches or convexities as sediment is stored within the channel.

Sediment discharge during the channel incision is characterized by a logarithmic decrease (Fig. 5-24) owing to a decreased rate of nickpoint migration, to a decrease of bank erosion, and to increased sediment storage in the channel. Considerable variability of sediment discharge occurred during the experiments reflecting bank caving and development of secondary nickpoints.

An additional series of experiments were performed by lowering base level again (Fig. 5-25). Therefore, the initial profile is the same as the 840-minute profile (Figure 5-23). Note that at the end of this series of runs, at 1,800 minutes, aggradation was general along the profile, and as compared to the 720 and 930-minute profiles the final profiles showed fewer irregularities. However sediment deposition produced the two convexities in the lower 5 m of the 1,800-minute profile.

As in the previous experiment, degradation was variable in space and time, and "it depended to a large extent on the profile irregularities that existed in the previous profile in the sequence" (Begin, 1979).

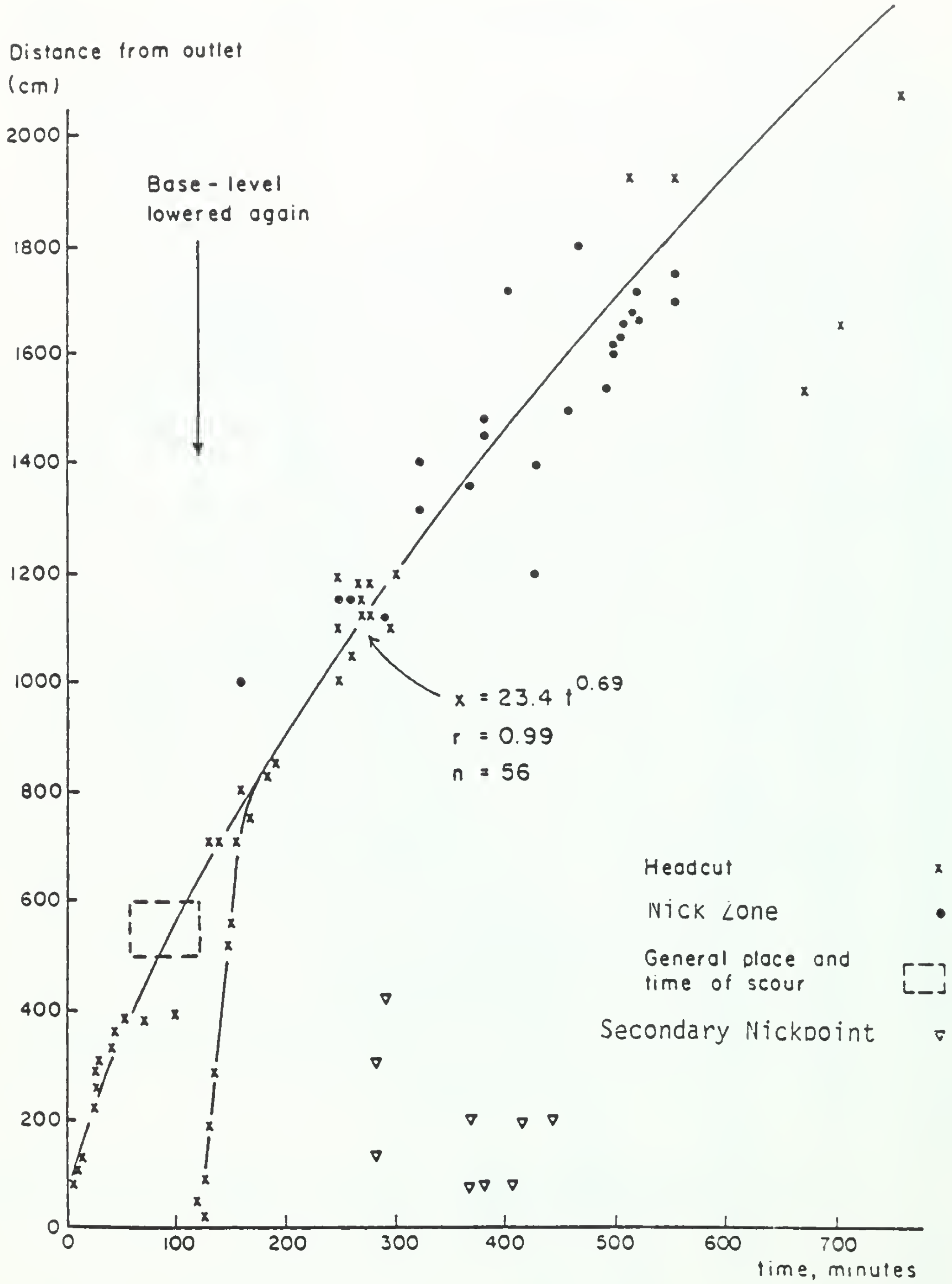

Figure 5-23. Nickpoint migration, Runs 3A and 4. t_0 = 42.00 hrs. (Begin, 1979)

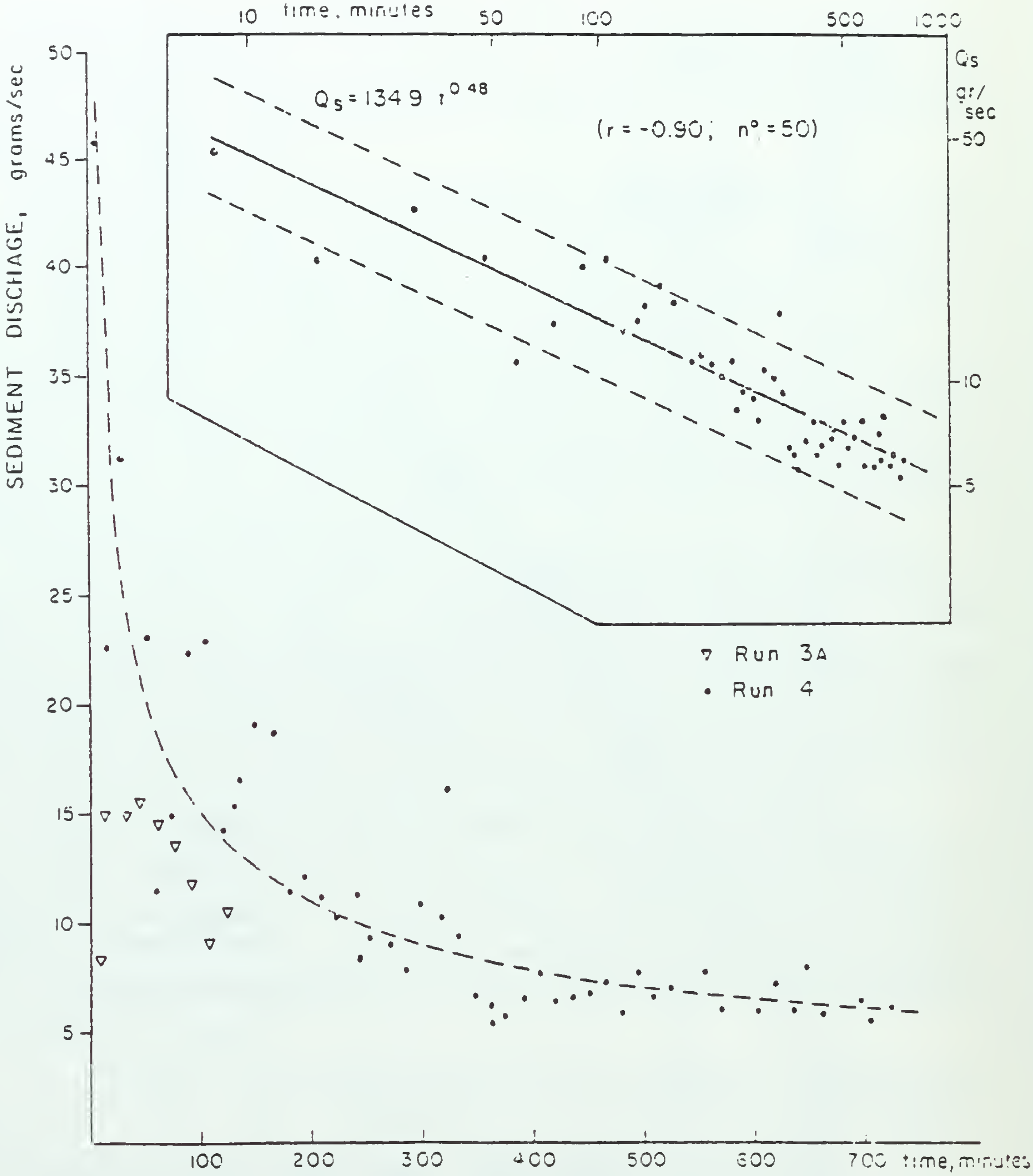

Figure 5-24. Changes with time of sediment discharge at the outlet, Runs 3A and 4; t_o = 42 hrs. for Run 3A, t_o = 44.00 hrs. for Run 4. (Begin, 1979)

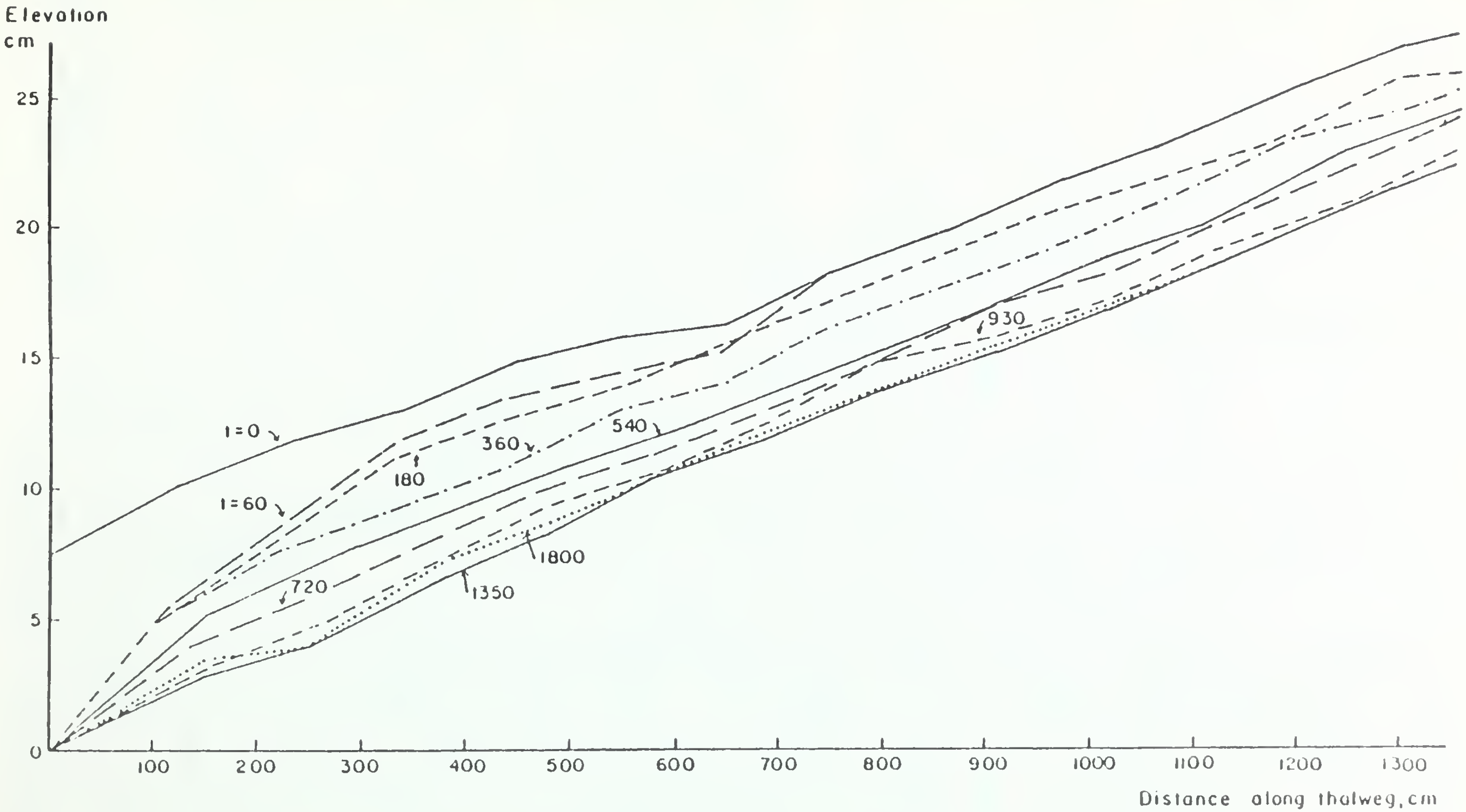

Figure 5-25. Profile development during Run 5. t_0 = 56.00 hrs. Numbers denote time in minutes. Profiles are based on mean bed elevations. (Begin, 1979)

The development of secondary nickpoints was a dominant feature of the run. These small headcuts, about 25 mm high, migrated rapidly and disappeared after moving between one and three meters. They formed in channel convexities, caused by local accumulation of sediment.

A third series of experiments on the effect of base-level lowering were carried out in a large flume. The sediments contained 20% more silt and clay than the previous experiments in the REF and they also contained about 4% more sediment that was coarser than sand size.

The increased cohesion resulted in maintenance of a nickzone following base-level lowering (Fig. 5-26), and the coarser fraction of the sediment developed channel armor that effectively prevented additional degradation. The result was a steeper, final profile.

The armoring also dramatically affected sediment discharge (Fig. 5-27). Sediment discharge decreased to about 7 gm/sec. and then the decrease was abrupt at 200 minutes although complete armoring of the channel was developed until after 360 minutes of run.

Channel widening was an important source of sediment, and of course, as degradation at a given location slowed, the channel widened in an effort to achieve a degree of stability. The development of alternate bars or point bars enhanced the widening process and this was the initial step in the eventual development of a sinuous inner channel (see discussion of Rio Puerco, Big Pine Creek, and Oaklimiter Creek).

Channel widening during the experiments was a function of amount of base-level lowering and amount of sediment in transport. The amount of base-level lowering affected the amount of widening in two ways. First, a large drop in base level produced high banks, that were subject to failure. The second effect of base-level lowering was on the amount of sediment produced by the migration of the nickpoint. Large drops in base level produced proportionately large amounts of sediment from nickpoint migration. The sediment produced by nickpoint migration accumulated as alternate bars which deflected the flow, thereby causing additional bank erosion and channel widening.

The amount of sediment transported was a contributor to the widening process. As long as sediment was produced by nickpoint migration, point bar building and bank failure occurred.

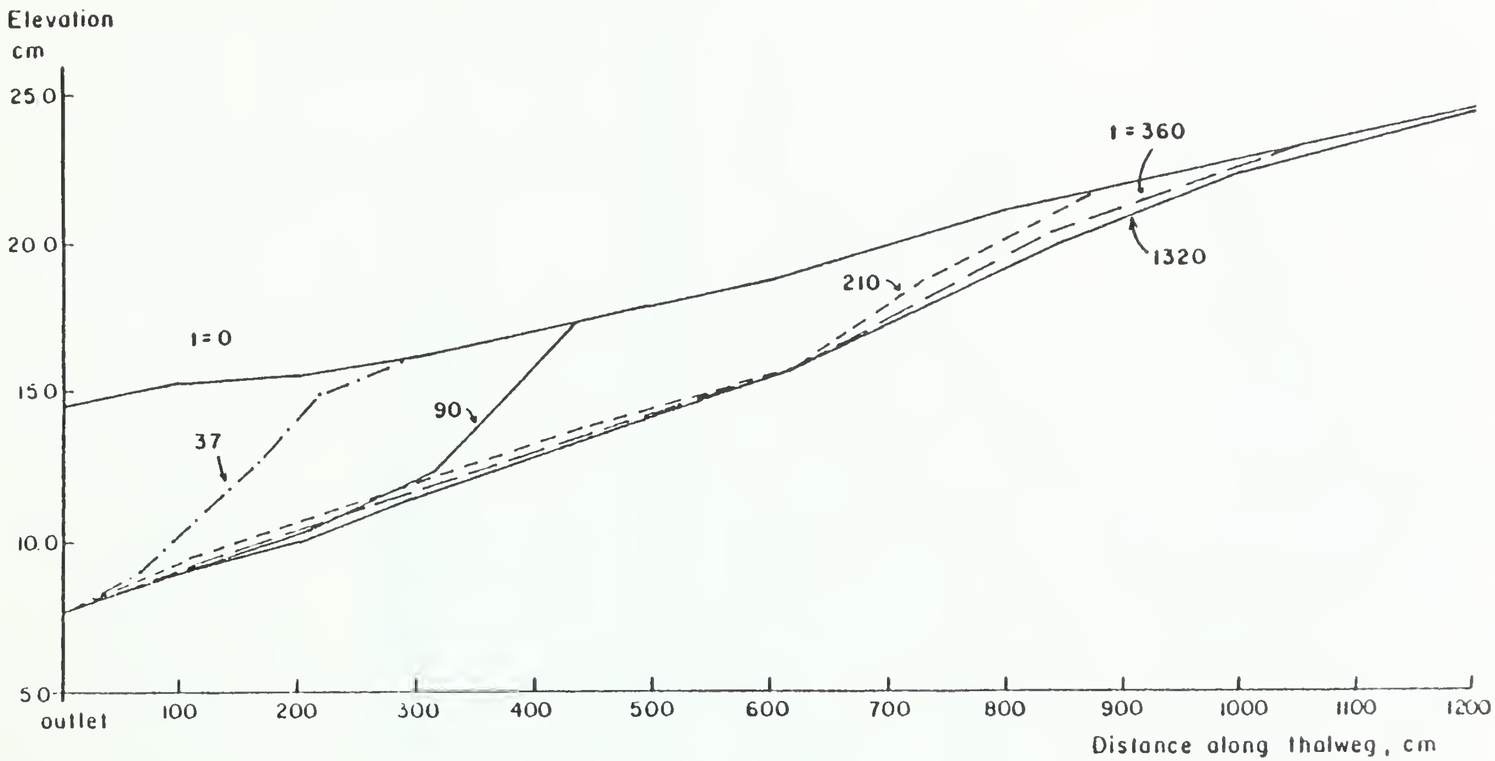

Figure 5-26. Profile development, Run 11. Numbers denote time in minutes. Profiles are based on mean bed elevation. (Begin, 1979)

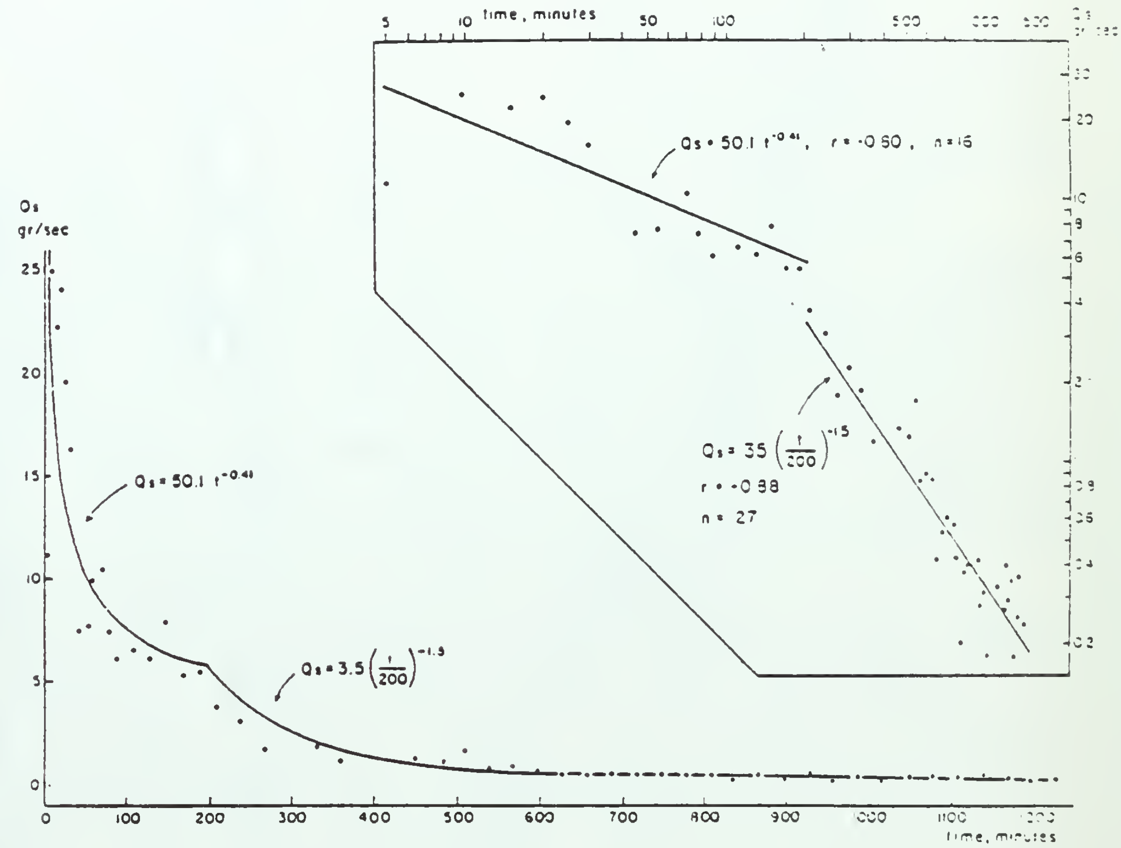

Figure 5-27. Changes with time of sediment discharge at the outlet, Run 11. (Begin, 1979)

Once the nickpoint stabilized, widening ceased and point bars were eroded away. However, when sediment was artificially fed into the channel, point bars formed again and the channel widened.

Cross sections were observed to go through phases of degradation, aggradation, and stability in the vertical direction and undercutting, accretion, and stability in the lateral direction (Fig. 5-28). Attempts at defining empirical relationships using the parameters computed from the cross section, longitudinal profile and discharge measurements which might explain or differentiate among these phases met with little success.

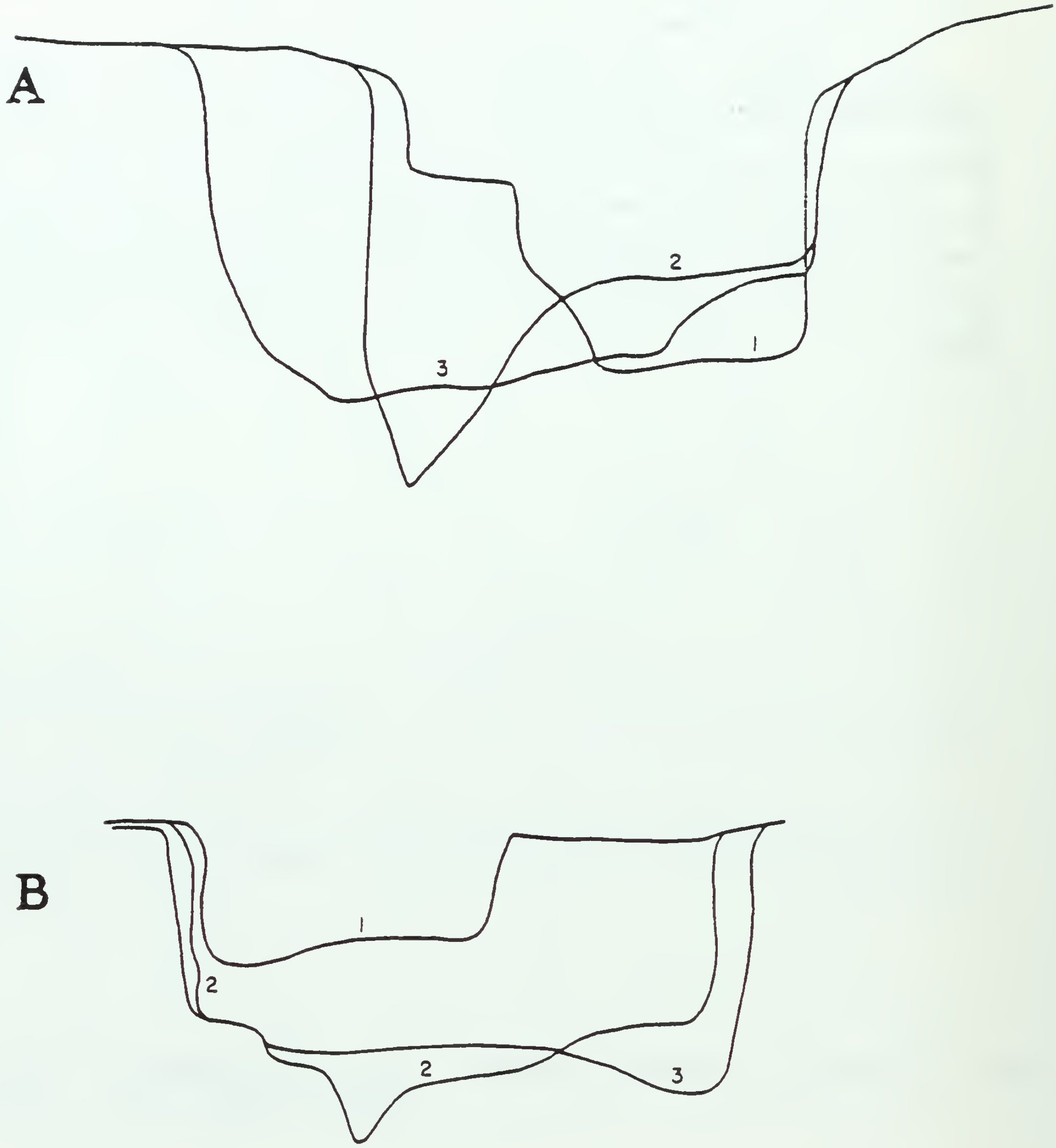

Figure 5-28. Series of cross sections at 2 m (A) and 1 m (B) from end of flume. Note that the oldest profile (3) is shallower than profile 2 and the alternate bar development on the oldest profile (Meyer, 1980).

6. MECHANICAL ASPECTS OF INCISED CHANNEL DEVELOPMENT

Studies of the soil mechanics aspects of incised channel development have been conducted by numerous investigators (Piest et al, 1973, 1975; Lohnes and Hardy, 1968; Bradford and Piest, 1977, 1980; Lohnes et al, 1980; Piest and Spomer, 1968; Thorne et al, 1981; Little et al, 1982). In general these studies have focused jointly on the problems of channel degradation and the resulting bank instability either in loess (Lohnes and Handy, 1968; Bradford and Piest, 1977, 1980; Lohnes et al, 1980) or valley-fill alluvium (Thorne et al, 1981; Little et al, 1982).

BANK STABILITY ANALYSIS

Lohnes and Handy (1968) identified two major mechanisms of failure of gully banks that were formed in loess; (1) shear failure along a planar slip surface through the toe of the bank, and (2) slab failure of vertical banks by tension cracking and plane slip. They investigated the relationships between the height, slope angle and physical properties (i.e. strengths) of the banks and then used a Culmann method (Lambe and Whitman, 1979) to analyze the shear failure and to predict maximum stable bank heights for non-vertical slopes. Slab failure was also analyzed with respect to vertical banks by the use of a modified Culmann method which took tension cracking of the upper bank into account. Their analysis can only be regarded as a approximation since the affects of pore pressures were neglected (Thorne et al, 1981). The Lohnes and Handy (1968) approach should, therefore, only be utilized when soils are highly permeable and when they have a low degree of saturation.

Bradford and Piest (1977) conducted a field investigation of the effects of pore-water pressure on bank failure in loess-derived alluvium. They observed that failure took place by a "pop out" near the toe of the bank, which left the upper bank in a overhanging state. The upper bank subsequently fails in shear. The "pop out" failure of the toe slope was probably due to seepage forces (Thorne et al, 1981). However, Bradford and Piest (1977) considered that seepage forces had little effect on the mode of bank failure.

Bradford and Piest (1980) identified three major mechanisms of gully wall failure: (1) deep seated circular arc toe failure, (2) slab failure, and (3) "pop out" failure with shear failure of the remaining cantilevered bank section. They related the occurrence of the failure types to gully development, and suggested that the slab failures were generally located just downstream of the headcut whereas circular arc failures were located further downstream. Slab failures were associated with vertical banks whereas circular arc failures were associated with lower angle banks. Very significantly, Bradford and Piest (1980) stated that fluvial erosion of intact bank material appears to contribute very little to bank retreat. Similar observations have been made by Little et al (1982) and Schumm et al (1981), and they reinforce the idea that most gully bank failures are essentially due to gravitational forces which are primarily controlled by the degree of channel incision. In contrast to the above statement, fluvial erosion of previously failed debris does play a significant role in determining the rates of bank retreat. This is because the fluvial activity controls the state of "basal endpoint control" (Thorne, 1981). Removal of the failed material results in the formation of steeper banks, and it may induce toe erosion. These factors lead to renewed bank erosion by mass failure, without basal flow, mass failure of the bank material would lead to bank slope reduction and stabilization within a relatively short period of time (Lohnes and Handy, 1968; Brunsden and Kesel, 1973; Thorne, 1981).

Thorne (1981) summarized the mechanisms that are associated with failures of stream banks. He considered that rotational types of failures are typical of cohesive bank materials that have great height and comparatively low slopes because the orientation of the principle stresses changes with depth. In contrast, for near vertical banks the failure surface is almost planar and the development of tensile stresses in the upper part of the bank generate vertical tension cracks which promote slab failure. The presence of tension cracks significantly effects the mechanism of failure and therefore the limiting bank height. Bradford and Piest (1980) demonstrated that the presence of tension cracks whose depths may be equal to half the bank height (Terzaghi, 1943), effectively reduced the limiting bank height by 50% when they utilized Lohnes and Handy's (1968) analysis for slab failures. Therefore, tension cracking plays an important role in controlling the stability, mode of failure and limiting bank height of steep banks.

Thorne et al (1981) and Little et al (1982) investigated bank failures along an incised channel in north-central Mississippi. They collected field data to describe critical bank heights and slope angles. They concluded that the dominant mode of bank failure was a slab failure in which tension cracking was an important component. Failure was observed to occur along a slip surface between the toe of the slope and the bottom of the tension crack. Development of tension cracks appeared to be related to repeated wetting and drying of the materials which have a high montmorillonite content (Harvey et al, 1983). Thorne et al (1981) used a log-spiral failure surface to model bank failure. By establishing values of bank height and slope angle they were able to define a line of critical stability for a given bank section (Fig. 6-1).

Figure 6-1 provides a means of integrating the studies of channel evolution following destabilization with the mechanisms of bank failure. Bank failure, and therefore, channel widening can be reduced by either of two mechanisms: (1) reduction of bank height and (2) reduction of slope angle. Both of these two mechanisms are observed to occur in studies of incised channel evolution. Bank heights are effectively reduced by deposition of sediment in the bed of the channel at some distance downstream of the headcut. However, this process can only occur when channel width is such that the reduction of the shear stresses prevents the removal of failed bank materials (i.e. basal endpoint control). This in turn will permit the bank angle to be reduced in a relatively short period of time (Thorne, 1981).

SOIL PIPING

A soil particle can only move in a pore space that is larger than itself and therefore soil piping can only take place where there are large voids in the soil. Such large voids are always present at a free face and they may also be present at the interface between a fine- and a

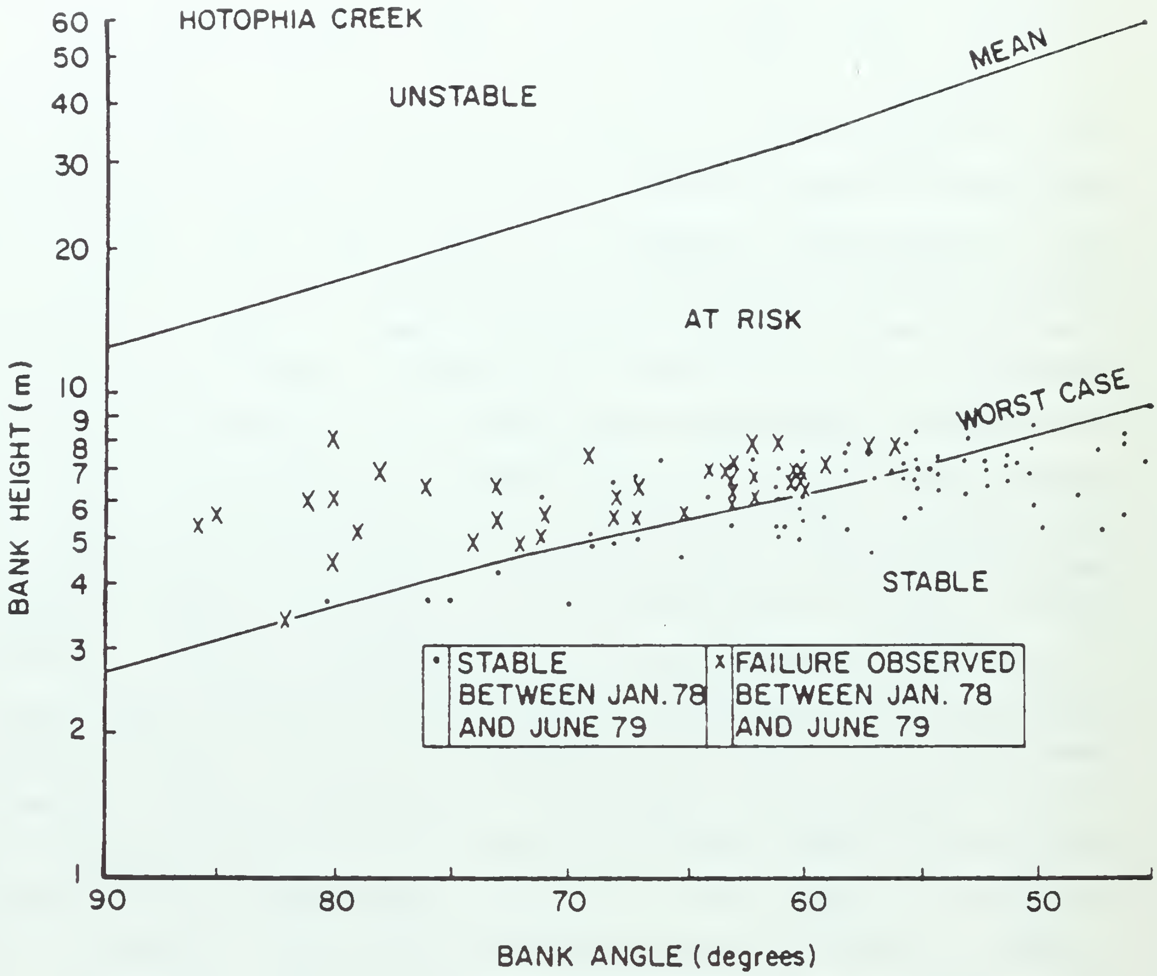

Figure 6-1. Bank Stability Graph for Hotophia Creek.

coarse-grained deposit. If flow goes across a free face or from the fine- into the coarse-grained deposit, then large soil grains may be moved which results in the destruction of the soil framework and the development of a new large void. Repetition of this process leads to headward erosion of continuous channels within the soil mass. Both of the above conditions for the presence of large voids are commonly present in alluvial deposits and as such these deposits are frequently piped (Carson and Kirkby, 1972). Pipes vary in size from a few millimeters up to several meters in diameter and they generally follow hydraulic gradients.

Gully development is generally regarded as being the result of surface erosion. However, a second alternative has received more attention in the most recent literature. It involves subsurface erosion by pipeflow and the ultimate collapse of the pipes to form gullies.

Jones (1971) has discussed the factors favoring the occurrence of pipes as, (1) susceptibility to cracking in dry periods, high silt clay content and high percentage of montmorillonite, (2) periodic high intensity rainfall and devegetation, (3) biotic breakup of soil and a relatively impermeable basal horizon, (4) an erodible layer above this base, high exchangeable sodium and a high base exchange capacity high in soluble salts, and, (5) steep hydraulic gradients. Heede (1971) emphasized the importance of steep hydraulic gradient, high exchangeable sodium content, low gypsum content and montmorillonite in the initiation of piping. Sherard and Decker (1977) concluded that piping was most likely to occur in alluvial, colluvial, lacustrine and loess deposits that were derived from lithologies of marine origin.

Laffan and Cutler (1977) and Drew (1982) have developed models for the development of gully systems that are originated by pipe formation. Imeson et al (1982) have suggested that gullies that are developed from piping have a characteristic "U" shape. This is in contrast to gullies developed by surface flow processes which were reported to have a "V" shaped morphology. However, Harvey (1982) considers that the main morphometric characteristic that is related to piping is that of channel alignment. He further notes that where erosional development has progressed to form a badland topography, no morphometric distinctions can be made between areas that have developed by the collapse of pipes and those that developed as a result of surface processes.

This review of the literature concerning piping and the development of gullies indicates that piping is a significant factor in the formation of gullies. This appears to be especially true when the following conditions are present: (1) steep hydraulic gradient, (2) dispersible soils, and (3) the materials are either alluvial or colluvial in origin.

7. SUMMARY AND CONCLUSIONS

Gullies are one member of a continuum of incised channels that range in size from rills to major incised channels (arroyos) that may be on the order of 10 to 15 meters deep. They are clearly major sediment producers (Patton, 1973). They cause major problems in many agricultural and rangeland areas (Bennett, 1939; Heede, 1974) and they may also pose hazards to man-made structures (Shen and Schumm, 1981). However, the presence of gullies does not necessarily indicate that sediment yields are high (Thornes, 1984). Although gullies have been investigated in many areas, general conclusions about their initiation, dynamics and control have rarely been forthcoming because climatic, soils and vegetational conditions have been very different (Schumm et al, 1984).

One of the major problems that has hindered the generation of general conclusions from the vast amount of research that has been conducted in incised channels, has been that of terminology. In an attempt to rectify this problem Schumm et al (1984) have proposed a classification (Table 1-1), and this classification has been utilized in this report.

The basic causes of incision are well known in that incision occurs when the eroding forces exerted by concentrated flowing water exceed the resistance of the materials subjected to the flow. The obvious causes of incised channel development are listed on Table 2-1. Some of the principle causes of incised channels of various orders are given in Table 2-2.

The primary objective of this report was to review the extensive literature dealing with incised channels of various orders and to develop a conceptual framework that can be used to formulate an easily applied technique which can be used to analyze:

(1) Threshold conditions for incised channel development,

(2) The stages of gully evolution,

(3) Current stability of a gully system,

(4) Long-term gully erosion rates and sediment yields. Each of these sub-objectives will be discussed and a conceptual framework will be developed in the remainder of this chapter.

THRESHOLD CONDITIONS FOR INCISED CHANNEL DEVELOPMENT

Incised channels cannot be regarded in isolation from the remainder of the fluvial system, and in fact they can be regarded as merely one component of drainage network development. McLane's (1978) experimental study of drainage network development concluded that network development is hydrophilic, and therefore, the extension of the drainage network should be predictable from analysis of pre-incision topography. Schumm (1973, 1977) has summarized the essence of a geomorphic approach to understanding and therefore predicting incised channel development:

(1) The land surface is complex, dynamic, and it changes with time.

(2) The land surface may respond dramatically during a short time due to the exceeding of a geomorphic threshold.

(3) The response of a complex landform to change is itself complex. That is, secondary responses will complicate the adjustment of the system to change.

Schumm and Hadley (1957) proposed one of the first models of gully prediction when they concluded that a cycle of erosion and deposition was associated with gullying in the arid and semi-arid western U.S. They suggested that as a result of sediment deposition on alluvial valley floors, the increased valley slope eventually caused an intrinsic threshold of stability to be exceeded, which in turn resulted in gullying. However, they also recognized the role of flood events of different magnitudes in determining the stability of the valley floors (Fig. 2-2).

Patton and Schumm (1975) developed a procedure whereby a threshold valley slope could be determined for the Piceance Basin of western Colorado, that would discriminate between gullied and ungullied valley floors (Fig. 2-1). In the absence of discharge data, Patton and Schumm (1975) used drainage area, (an index of discharge; Burkham, 1966), and valley slope to define the discriminant function (K). They concluded that this parameter did not bear a clear relationship to hydraulic variables such as shear stress or stream power, and that the distance of a point from that line was but a qualitative measure of valley instability.

Begin and Schumm (1979) re-evaluated Patton's (1973) data and derived a threshold parameter that was based on the shear stress exerted by the flow on the valley floor. This relative shear-stress indicator (Equation 4-7), could be used to discriminate between gullied and ungullied valley floors. However, Begin and Schumm (1979) emphasize that increased values of the relative shear-stress indicator do not always imply that gullying will take place. Rather, it implies only that there is an increase in the probability that the valley floor will be gullied. Their caution is reasonable because of the stochastic mechanisms involved, such as the temporal and spatial distribution of rainfall and the nonhomogeneity of soil and vegetation distribution.

Bradley (1980) utilized valley floor slope, valley width, and drainage area to derive a threshold zone of valley floor instability in northeastern Colorado (Fig. 4-15), which was based on a critical-slope/critical-width ratio. However, all three variables of gully initiation were considered when the threshold conditions were defined in terms of relative stream power (Equation 4-16). There are two means by which the stream power geomorphic threshold can be exceeded and gully erosion initiated. This is achieved by either an increase in discharge or by an increase of valley-floor slope.

In summary, it can be concluded that the recognition and quantification of the variables described above that identify the erosional geomorphic thresholds at which gullies are initiated provides an acceptable means of predicting the stability of alluvial valley floors. It is important to note that this approach is purely deterministic and the stochastic behavior of rainstorm events which trigger gullying requires a probabilistic approach (Fig. 4-20). The relations derived by Begin and Schumm (1979) and Bradley (1980) can also be used to evaluate the effects of land use changes within a drainage basin (Figs. 4-11, 4-15). Providing that the effects of the land use change on the rainfall-runoff relationship can be quantified, the resultant discharge increase, in terms of basin area, can be located along the abscissa of Figs. 4-11 and 4-15. Thus, the effect of increased discharge on the stability of a valley reach can be ascertained.

It may be concluded that the threshold conditions are not fixed and any land use, hydrologic or climatic alteration can change them. If the changes affect the entire area the threshold line or zone may shift downward, which will render additional valley floors unstable.

When the change is local, it may have the effect of shifting a plotted point to the right, thereby placing it closer to or across the threshold line. Theoretically there should be no points that plot a considerable distance above the threshold line, but in fact, many valleys do. This may indicate that the threshold has shifted downward. Increased agricultural activity and the drought years of the 1930's plus extreme storm events have the capability of causing this downward shift of the threshold line.

The obvious deduction is that within any area, erosional activity will be variable and at least the smaller components of the landscape need not be in the same stage of erosional or depositional development. For the most part this is exactly tne case encountered in semi-arid regions of western United States. However, the relations developed for the Piceance Creek and Chalk Bluff areas indicate that the distribution of erosional features can be explained and, therefore, the sites of additional erosion may be predictable.

THE STAGES OF GULLY EVOLUTION

In spite of the apparent diversity of incised channel morphology, which is related to climatic conditions and to the character of the materials into which the channel is incised, all incised channels follow essentially the same evolutionary trend with eventual development of relatively stable conditions. The evolutionary development of the incised channels can be interrupted by changes in environmental conditions, such as further base level lowering or increased discharge resulting from changes in land use. The evolutionary development occurs in all orders of incised channels, and the evolution will normally be episodic and complex as the histories of Layton's and Walden's gullies reveal (Fig. 4-5).

Ireland et al (1939) described a four-stage gully evolution model for valley-side gullies which ultimately resulted in a period of stabilization that was characterized by the slow development and accumulation of new topsoil over the old scarred surface. Bariss (1977) showed that the slopes of the gully sidewalls reduced from greater than 45^{o} to a mean value of 36^{o} when the gully had stabilized. It is evident from these two investigations that valley-side gullies evolve to a condition of relative stability when the channel has cut to a new base level and the gully walls have reclined and have been vegetated.

Schumm and Hadley (1957) described a cycle of gully erosion in semi-arid regions that involved incision of the valley floor and the eventual refilling of the incision to its former elevation. Heede (1967) commented that discontinuous valley-floor gullies are representative of the most youthful stage of valley instability that eventually leads to the formation of continuous valley-floor gullies. Brice (1966) described the erosional evolution of valley-floor gullies, and he was able to relate the locations of headcuts and gullies to characteristics of the valley floor. The cited studies indicate that through time valley-floor gullies evolve from a condition of disequilibrium to that of a new state of equilibrium in much the same manner that valley-side gullies evolve.

Entrenched streams (i.e. arroyos) of the southwest appear to evolve in similar fashions. Thornthwaite et al (1942) concluded from their study of Polacca Wash that there were stages of arroyo evolution as follows: (1) initiation; (2) enlargement by headward elongation; (3) healing, by reduction of slope of the walls and establishment of vegetation; and (4) stabilization, revegetation and possible eventual filling or obliteration. However, unlike the characteristic decline of gully walls in more humid regions, lateral cutting maintains steep valley walls, which tends to enlarge the arroyo. Elliott (1979) reported that the lower reaches of Rio Puerco had evolved from Type 1 to Type 2 reaches over a period of 100 years (Fig. 5-6).

An important distinction should be made between aggradation in small and large ephemeral stream drainages. Aggradation in discontinuous gullies and small arroyos may often result in complete filling of the channel as the cycle of erosion progresses (Schumm and Hadley, 1959).

However, in large arroyos, such as Rio Puerco, the magnitude of the runoff events is such that they tend to prevent the channel from aggrading to its former level. In this respect, channels flowing through large arroyos behave more like alluvial rivers in valleys than incised streams (Patton, 1973).

Schumm (1961) and Patton and Schumm (1981) demonstrated that the nature of sediments that were being transported through arroyos significantly affected the morphology of the channel and the nature of the channel behavior. Schumm (1977) incorporated the nature of the sediment loads into a classification of alluvial channels (Table 5-2). Watson and Harvey (1984) used the data of Schumm (1961) and Patton and Schumm (1981) to derive a multi-variate relationship between the equilibrium channel top width of these incised channels and an energy parameter (AGI), d_{50} of the bed sediments and the percentage of silt and clay (M) in the bed and banks of the channel (Equation 5-4). The relationship was predicated on the assumption that the channels went through and evolutionary sequence of development and that the new equilibrium state could be recognized.

Studies of channelized streams (Daniels, 1960; Barnard, 1977; Harvey et al, 1983) indicate that these channels have responded to man-induced incision in a very similar manner to the other forms of incised channels (Fig. 5-18). In contrast to the lower order incised channels (e.g., valley-side gullies, valley-floor gullies, small arroyos), the channelized streams tend to evolve to a condition where a new floodplain is formed within the entrenchment, and therefore, they behave in a similar manner to the large arroyos. Barnard (1977) indicated that the equilibrium slope of an incised channel would approximate the pre-channelization slope. This conclusion is reasonable since valley slope is an important independent variable that controls channel slope (Harvey, 1983). In Oaklimiter Creek, Harvey et al (1983) were able to quantify the new equilibrium state of the channel as follows: (1) channel slope reduced to about 0.001; (2) width-depth ratio (F) increased to 6; 2.5 to 3 feet of sediment accumulated in the bed of the channel; and (4) development of a meandering thalweg at low-flow stages.

The experimental studies of channel incision (Begin, 1979; Begin et al, 1980, 1981) provided a detailed record of channel incision following base level lowering, and they essentially confirmed the conclusions reached from field observations of arroyos, channelized streams and gullies. The adjustment of a channel to a base-level lowering or channelization, in all cases is found to be complex and episodic with periods of incision followed by sediment storage and in some cases renewed incision.

CURRENT STABILITY OF A GULLY SYSTEM

Studies of incised channel evolution have indicated that following incision a progressive sequence of development occurs, and the end member of the sequence is a stable or equilibrium channel form. Schumm et al (1981) and Watson and Harvey (1984) developed a model of channel evolution that was based on a location-for-time substitution technique (Paine, 1984 in press). The use of a channel evolution model to assess the current stability of a gully system depends on the ability to determine the new equilibrium state of the channel. If this can be achieved then relationships of the type derived by Harvey et al (1983) (Equation 5-6) and Watson and Harvey (1984) (Equation 5-4) can be utilized to quantify the equilibrium top width of the gully. Spatial interpolation techniques (Park, 1977) can then be used to project the equilibrium form into non-equilibrium reaches of the gully, thereby enabling an estimate to be made regarding the future adjustment of the system (Harvey et al, 1983).

Foster (1964) referred to the equilibrium grade of gullies as a silting grade, but he noted that due to differences in soils, there is no exact grade that is stable under all conditions. Probably the best and easiest way to determine a silting grade is to investigate the grade of gullies in the area that appear to be stable, that is, that are not actively eroding in the bottom of the channel. Barnard (1977) indicated that the equilibrium grade of Big Pine Ditch would approximate that of the pre-channelized stream. Harvey and Watson (1984) used Equation 5-6 to predict an equilibrium channel slope. Since the relationship was derived for equilibrium reaches then it is possible to manipulate the relationship to determine an equilibrium channel slope. Area-Gradient Index (AGI) is the product of the drainage area (DA) and channel slope (S) which can be substituted in equation 5-6 as follows:

$$TW = A(AGI)^b = a(DA \times S)^b.$$

$$S = \frac{\left[\frac{TW}{a}\right]^{1/b}}{DA} \qquad (7\text{-}1)$$

From a soil mechanics point of view, Thorne et al (1981) demonstrated that critical bank stability criteria could be defined by a relationship between bank slope angle and bank height (Fig. 6-1). Bariss (1977) showed that stable gullies had side slopes that were on the order of 36^o, which had been reduced from greater than 45^o when the gully was in an unstable form. Watson and Harvey (1985, in preparation) have related channel top width to the change in channel depth. Channel depth is controlled by the amount of incision that is required to reduce the channel slope to an equilibrium value (Equation 5-5).

Recognition of a current condition of stability of a gully system is dependent on the ability to define equilibrium reaches of the channel. Because gullies are found to occur in many different soil and climatic environments it is not possible to define general conditions for the recognition of gully stability. However, studies of a range of gullies in a specific area can provide criteria to identify conditions of stability. These criteria would include, channel slope, valley slope, bank heights (i.e. channel depths), bank angles, and channel top widths. In the absence of discharge data to quantify the energy in the system, drainage area or a parameter such as AGI can be used.

LONG-TERM GULLY EROSION RATES AND SEDIMENT YIELDS

Prediction of long term gully erosion rates and sediment yields is hampered by the stochastic nature of the distribution of precipitation events. The occurrence of intense rainfall events covering relatively small areas is a distinctive feature of thunderstorm precipitation in the western United States (Leopold, 1951). Also, higher intensity thunderstorms are generally more limited in areal distribution. Further, a spatial and temporal problem is introduced into the prediction of sediment yields when the size of the drainage basin is considered. Boyce (1975) showed that sediment delivery ratio is inversely related to drainage basin area and his results are confirmed by the experimental work of Parker (1977), McLane (1978) and Begin (1979). The experimental studies showed that sediment yield decreased logarithmically with time because as the nickpoints migrated into areas of reduced drainage area (i.e. discharge), the rate of nickpoint migration was also reduced. The effect of reduced nickpoint migration rate was compounded by the fact that the drainage network had expanded to the point where storage of sediment, derived from nickpoint migration, was also occurring.

Schumm and Parker (1973) and Schumm (1977) demonstrated that sediment yields from an experimental drainage basin varied with time and they attributed this to the complex response of the system, whereby sediment was repeatedly stored and flushed from the system (Fig. 3-8). On a larger scale Womack and Schumm (1977) demonstrated episodic behavior of Douglas Creek. Repeated cycles of cut and fill (Fig. 2-4) would have resulted in variations in sediment yield through time even though the general trend in Douglas Creek was one of incision. On a smaller scale, their observations in Douglas Creek were supported by the work of Ireland et al (1939) in Layton's and Walden's gullies, where convexities on the longitudinal profiles indicate that sediment was being stored and flushed within the system (Fig. 4-5). The concept of geomorphic thresholds (Schumm, 1973, 1977) also applies to the temporary storage of sediment and development of the convexities on the longitudinal profiles. The increased gradient will eventually lead to incision and removal of the sediment and the convexity. Although threshold parameters have been identified for the initiation of gullying, (Patton and Schumm, 1975; Begin and Schumm, 1979; Bradley, 1980), it is unlikely that the same threshold conditions will be applicable to the incision of the secondary deposits since the physical condition of the deposit has changed and flow is now channelized.

The complexities involved in the determination of gully erosion rates and sediment yields as enumerated above, suggest that the problem should be approached from two separate but obviously related viewpoints: (1) onsite, and (2) offsite. The onsite viewpoint deals mainly with sediment production by gullying and slope erosion. In essence the onsite viewpoint deals with changes in the landscape and the loss of a soil resource. The offsite viewpoint on the other hand deals mainly with the problem of sediment yields and ultimately with sediment delivery to reservoirs. Onsite and offsite are the two components of a sediment delivery ratio.

Onsite gully erosion rates tend to decrease exponentially with time because the rate of headcut migration decreases as the gully elongates into areas with lesser drainage areas and less discharge (McLane, 1978; Parker, 1977; Thompson, 1964). However, Begin (1979) has shown that erosion rates decrease exponentially even with constant discharge due to "diffusion" characteristics of the gully process.

Begin (1979), and Begin et al (1980, 1981) proposed the usage of the diffusion equation to predict the degradation process in response to base level lowering. This equation permits the amount of degradation (ΔY), relative to the amount of base level lowering (Y) to be determined at a time (t) after the initiation of the degradation process. For a two-dimensional analysis:

$$\frac{\Delta Y}{Y} = 1 - \mathrm{erf}\ (x/2\sqrt{(kt)}) \qquad (7\text{-}2)$$

in which erf is the error function, Y is the amount of base-level lowering, ΔY is the amount of degradation at time t after base-level lowering at a distance x from the channel mouth, and k is a diffusion constant, which is mainly dependent on soil properties and water discharge per unit width. An example calibration of k is shown in Fig. 7-1. Various manipulations of this equation can, therefore, be used to predict gully erosion rates.

The rates of nickpoint migration are only one of the factors that

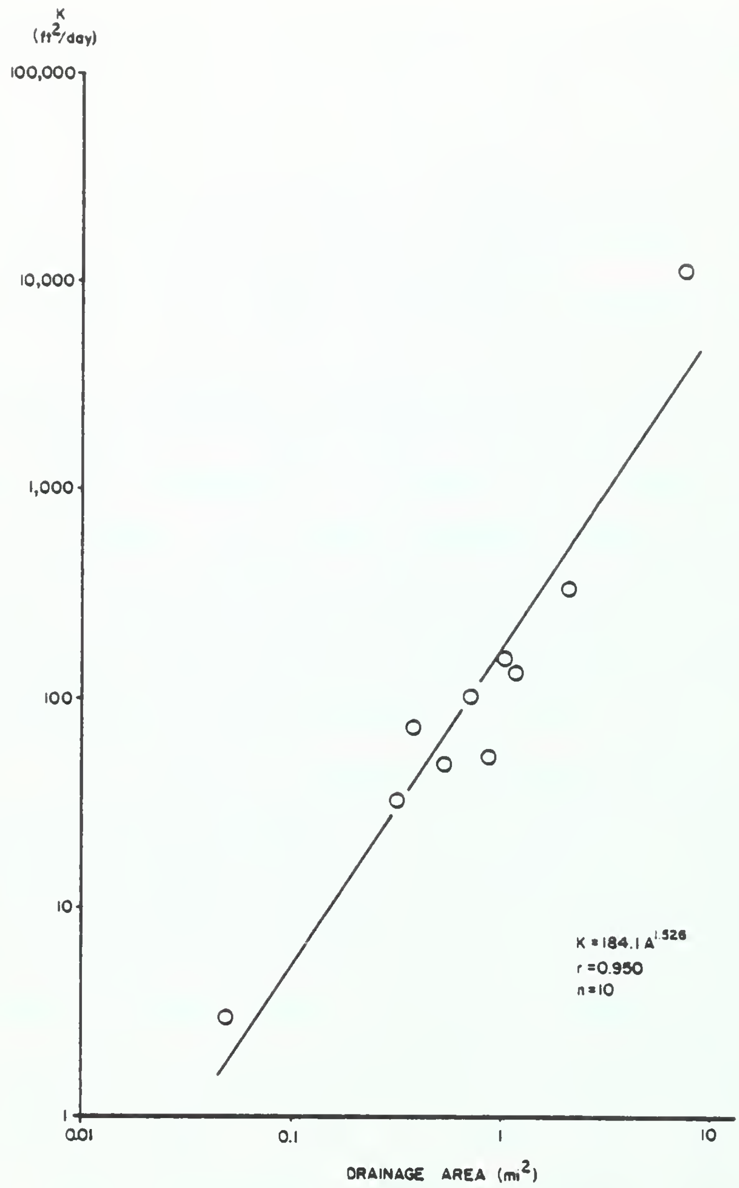

Fig. 7-1. The relationship between drainage area and the diffusion erosion coefficient k for 10 tributaries of the Oaklimiter Creek, Miss. Degradation of the tributaries occurred as a result of lowering of their base levels, ensuing from the degradation of Oaklimiter Creek. Such a relationship can be used in order to estimate k for other streams in that area, entrenched in similar sediments. (from Begin, in prep.)

must be determined for the quantification of onsite effects of gullying. Various studies of incised channels of all orders have shown that once incision has been initiated, then, the channel will go through an evolutionary sequence of development. In all cases this developmental sequence involves an increase in channel width which is predisposed by the increase in channel depth that is related to the degree of incision necessitated by slope adjustment. Harvey et al (1983) and Watson and Harvey (1984) have shown that the equilibrium top width can be predicted (Equations 5-6 and 5-4), and therefore comparison of existing top width for a reach of the channel with predicted equilibrium top width will provide an estimate of the expectable increase in channel width. Harvey et al (1983), related sediment yield (i.e. voiding of sediment) to increase in channel top width (Equation 5-7). Watson and Harvey (1984 in prep.) have demonstrated that channel top width can be predicted by channel depth (Equation 5-5). However, this approach will require calibration for the types of sediment that are encountered in different environments (Patton and Schumm, 1981).

When offsite effects of gully erosion are considered, estimates of sediment yield from gully/arroyo systems have been made by various authors (e.g., Bryan and Post, 1927), but Piest et al (1975) had a spectacular lack of success in using storm runoff to predict sediment yields from gully systems in Iowa. Measured sediment discharges ranged from 50 to 200 percent of the predicted values two out of three times. They emphasized the fact that the predicted values from one gully had even less value when applied to another gully. Sediment yields and sediment delivery ratios have been derived in general from studies of reservoir sedimentation (Strand, 1975; Boyce, 1975; Hadley and Schumm, 1961; Dendy and Champion, 1977). The use of reservoir sedimentation data to determine sediment yields and sediment delivery ratios presents some serious problems. The presence of a dam or reservoir raises local base level in a channel and this in turn causes deposition upstream of the reservoir and therefore reduces sediment delivery to the reservoir. Clarkin (pers. comm.) who is currently conducting an intensive study of sediment distribution downstream of active gully sites in eastern

Colorado considers that the sediment delivery ratio to a small reservoir is between 25 and 30 percent. In contrast, the USDA, Soil Conservation Service sediment delivery ratio - drainage area relationship predicts a sediment delivery ratio of between 50 and 53 percent (Strand, 1975). The reason for the disparity in sediment delivery ratios is that a large percentage of the sediment voided from the active gullies is accumulating in secondary storage sites, that are located upstream of the reservoir.

Without doubt, this sediment will be delivered to the reservoir in time, but at present it is not possible to predict the threshold conditions that will result in incision of the secondary sediment deposit. Further, there is no way of predicting at this time, whether incision of the secondary deposit will result in the accumulation of a tertiary deposit some distance upstream of the reservoir. This topic of secondary deposition and subsequent remobilization is an area that is overdue for some intensive field research.

In summary, it is apparent that onsite effects of gully erosion can be quantified. Gully migration rates and the ultimate gully configuration can be determined. The effects of gully erosion, namely sediment yields and sediment delivery ratios, are much more difficult to predict with any degree of precision and this area is therefore one which would benefit from further research. However, it is also important to recognize that storage of sediment, in smaller systems, is equated with stability (Schumm, 1961), and that efforts can be made to enhance storage of sediment.

This summary and the conclusions drawn from the extensive literature on incised channels indicate that the onsite effects of gullying can be more easily quantified than can the offsite effects. Although it is easier to quantify the onsite effects of gullying, it is not the intention of this report to suggest that the offsite affects cannot be quantified. Rather, the intention is to indicate that quantification of the offsite effects are considerably less precise because our knowledge of this aspect of system behavior is less than that for the onsite aspects.

A conceptual framework for the determination of onsite effects of gullying in a given area is presented in Table 7-1. The objectives of this framework are to provide a means to determine: (1) Conditions that define gully initiation; (2) Rates of gully extension; and (3) Land or soil resource loss as a result of gullying. The data that are needed to make these determinations are listed in Table 7-1. The data will have to be derived from field measurement and sampling, aerial photos and topographic maps. The appropriate analyses for the data are also shown in in Table 7-1. It is suggested that the relative shear-stress indicator and relative stream power indicator approaches should also be applied to the secondary and tertiary sedimentary deposits, and they should be compared to the shear stress and stream power threshold values for the incision of the valley floor.

A conceptual framework for the determination of offsite effects of gullying in given area is presented in Table 7-2. The objectives of this framework are to provide a means to determine: (1) Sediment yields; and (2) Sediment Delivery Ratios. The data required to meet these objectives and the required data analyses are also shown in Table 7-2. Sediment Delivery Ratio will have to be corrected for the volume of sediment in secondary or tertiary storage sites.

Finally, it should be emphasized that any relationships that are derived from studies of gullies in a given area cannot be readily extrapolated to other areas where the environmental setting (i.e. soils, vegetation, topography, geology) may be very different. Also, recognition of threshold conditions permits identification of incipiently unstable components of landscapes permitting a preventive conservation approach to the problem of semi-arid erosion.

Table 7-1. Conceptual framework for the determination of onsite effects of gullying.

OBJECTIVES:

Determination of: (1) Conditions for Gully Initiation
(2) Rates of Gully Extension
(3) Land/Soil Resource Loss

DATA REQUIREMENTS AND SOURCE OF DATA*

(1) Valley slope, valley width (F)
(2) Gully bed slope (F)
(3) Gully length (F)
(4) Gully widths (F)
(5) Bank angles (F)
(6) Drainage area (T)
(7) Bed material gradation (F)
(8) Percent silt and clay in perimeter sediments (F)
(9) Date of gully initiation (P)

ANALYSES

(1) Relative shear stress threshold } valley floor and	(Equation 4-7)
(2) Relative stream power threshold } secondary deposits	(Equation 4-16)
(3) Top Width = $f[(AGI) \cdot d_{50} \cdot M]$	(Equation 5-4)
(4) Gully Slope = $f[TW/a]^{1/b}/DA$	(Equation 7-1)
(5) Top Width = $f(\Delta depth)$	(Equation 5-5)
(6) $\Delta Y/Y = 1 - erf(x/2\sqrt{(kt)})$	(Equation 7-2)
(7) Bank Height = f(bank angle)	(Equation 6-1)

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*F - Field measurement; T - Topographic map; P - Aerial Photographs

Table 7-2. Conceptual framework for the determination of offsite effects of gullying.

OBJECTIVES:

Determination of: (1) Sediment Yield
(2) Sediment Delivery Ratio

DATA REQUIREMENTS AND SOURCES OF DATA*

(1) Drainage area (T)
(2) Time of gully initiation (P)
(3) Voided volume of gully (F)
(4) Top width of gully (F)
(5) Volume of material in secondary and tertiary deposits (F)
(6) Reservoir sediment volumes (F)

ANALYSES

(1) Unit Sediment Yield = f(ΔT.W.)	(Equation 5-7)
(2) Sediment Yield = f(D.A.)	
(3) Sediment Yield = f(time)	(Figure 5-24)
(4) Sediment Delivery Ratio = f(D.A.)	

*F - Field Measurement; T - Topographic Map; P - Aerial Photographs

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